SWIPA Background Science
(Provisional material)

The SWIPA scientific assessment report that is currently under production is the formal background for the SWIPA assessment.

The following pages comprise a compilation of the (peer reviewed) authors' manuscripts of those parts of the SWIPA assessment that have been subject provisional technical/linguistic editing. The purpose of this document is to make these texts available to those interested in examining the source information behind the findings presented in the *SWIPA 2011 Executive Summary* document. Additional sections will be added as technical/linguistic editing proceeds.

The SWIPA scientific background report includes the following 12 chapters:

1. Introduction
2. Arctic Climate: Recent Variations
3. Climate Model Projections for the Arctic
4. Changing Snow Cover and its Impacts
5. Changing Permafrost and its Impacts
6. Changing Lake and River Ice Regimes: Trends, Effects and Implications
7. Mountain Glaciers and Ice Caps
8. The Greenland Ice Sheet in a Changing Climate
9. Sea Ice*
10. Arctic Societies, Cultures, and Peoples in a Changing Cryosphere*
11.
    11.1 Synthesis of Feedbacks and Interactions: From the cryosphere to the climate system – effects over various spatial and temporal scales
    11.2 Sea Level Change
    11.3 Contaminant Pathways and Change in the Cryosphere*
    11.4 Impacts of Changing Snow, Water, Ice and Permafrost on Arctic Ecosystems*
    11.5 Observational Needs and Knowledge Gaps for the Cruyosphere
12. SWIPA Synthesis: Implications and Findings*

*technical/linguistic editing still pending

Disclaimer: The material in this document may still contain errors. Every effort is being made to check the information presented in the SWIPA assessment at all stages of the report production process. However, until such time as the complete scientific assessment report has undergone editing, graphical production and layout, been subjected to final proofing by the authors, and been officially published, the report remains a provisional draft. Notwithstanding this, the SWIPA lead authors have formally confirmed that they are in full agreement with all statements in the *SWIPA 2011 Executive Summary* document as accurately and correctly reflecting the findings of the assessment that will be presented in the full SWIPA scientific background report.

Chapter 1. Introduction

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1.1. Why assess the effect of climate change in the Arctic cryosphere?

The *Arctic Climate Impact Assessment* (ACIA, 2005) was the second climate assessment conducted by the Arctic Monitoring and Assessment Programme (AMAP), one of the Arctic Council's expert Working Groups. It was produced in close cooperation with the International Arctic Science Committee (IASC) and the Arctic Council Working Group on Conservation of Arctic Flora and Fauna (CAFF). The ACIA report provided the first comprehensive documentation of the ongoing climate change within the Arctic and its potential impacts at local, regional and global levels, and formed an important background report for the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC, 2007). The ACIA report was the first Arctic climate impact assessment to evaluate and synthesize information on physical changes, changes in biological systems, and impacts on human infrastructure and activities.

The ACIA summary report (ACIA, 2004) identified ten key findings:

1. Arctic climate is now warming rapidly and much larger changes are projected.
2. Arctic warming and its consequences have worldwide implications.
3. Arctic vegetation zones are very likely to shift, causing wide-ranging impacts.
4. Animal species' diversity, ranges, and distribution will change.
5. Many coastal communities and facilities face increasing exposure to storms.
6. Reduced sea ice is very likely to increase marine transport and access to resources.
7. Thawing ground will disrupt transportation, buildings, and other infrastructure.
8. Indigenous communities are facing major economic and cultural impacts.
9. Elevated ultraviolet radiation levels will affect people, plants, and animals.
10. Multiple influences interact to cause impacts to people and ecosystems.

In its policy response to the ACIA, the Arctic Council (Arctic Council, 2004):

> *Acknowledge[d] the need to further organize the work of the Arctic Council and its subsidiary bodies based on the findings of the ACIA and direct[ed] the SAOs to report on the progress made at the 2006 Ministerial Meeting.*
>
> and
>
> *Direct[ed] relevant technical working groups of the Arctic Council to review the scientific chapters of the ACIA in the context of their ongoing and future work programmes and to report on the progress made at the 2006 Ministerial Meeting.*

New observations showed (and continue to show) that in recent years some of the components of the Arctic cryosphere have undergone changes that exceed even those described and projected in the ACIA (2005) and IPCC (2007) assessments. Consequently, the Arctic Council initiated an assessment of 'Climate Change and the Cryosphere: Snow, Water, Ice and Permafrost in the Arctic (SWIPA)' in 2008 (SWIPA project description, 2008).

The Arctic Council requested AMAP to undertake the assessment in collaboration with relevant international organizations (IASC and the International Arctic Social Sciences Association, IASSA) and relevant international activities (the World Climate Research Programme's Climate and Cryosphere Project, CliC; and the International Polar Year, IPY).

1.2. An Arctic cryosphere assessment

1.2.1. What is the Arctic?

The geographical delineation of the Arctic used by the SWIPA assessment is based on the definition used by AMAP (see Figure 1.new). The 'AMAP area' essentially includes the terrestrial and marine areas north of the Arctic Circle (66°32' N), and north of 62° N in Asia and 60° N in North America, modified to include the marine areas north of the Aleutian chain, Hudson Bay, and parts of the North Atlantic Ocean including the Labrador Sea.

1.2.2. What is the cryosphere?

The cryosphere collectively describes elements of the Earth System containing water in its seasonally and perennially frozen state. In the Arctic, the various components of the cryosphere include the following: snow, including solid precipitation; permafrost areas (i.e., ground that remains at or below 0 °C for two or more consecutive years) present in terrestrial and marine environments; river and lake ice; mountain glaciers and ice caps; the Greenland Ice Sheet; and sea ice in all its forms (i.e., perennial pack ice, seasonal land-fast ice). These cryospheric components (see Figure 1.1) represent a globally unique system, parts of which are inextricably linked with each other, with the landscapes, seascapes, ecosystems and humans in the Arctic, and with the global climate and ecological systems themselves. Consequently, shifts in the Arctic cryosphere have great significance, not just regionally within the Arctic but also globally across the planet as a whole.

Figure 1.1. [Arctic cryospheric components: either a graphic and/or series of photos of such.]

1.3. Aims of SWIPA

The objectives of the SWIPA Project are to provide the Arctic Council with timely, up-to-date, and synthesized scientific knowledge about the present status, processes, trends, and future consequences of changes in Arctic sea ice, melting of the Greenland Ice Sheet, and changes in Arctic snow cover, permafrost, mountain glaciers and ice caps, and related hydrological conditions in the Arctic. Future scenarios have been developed to determine, as far as possible, the consequences of these changes on physical processes on local, regional, and global scales, and to determine consequences for Arctic biological systems, and human societies and lifestyles.

The ACIA assessment constitutes a benchmark for the SWIPA assessment, which focuses on recent change in the Arctic cryosphere and the effects of observed and projected change. The SWIPA assessment is thus an update and extension to the ACIA findings on the consequences of change in the Arctic cryosphere component of the global climate system. Figure 1.2 presents a conceptual illustration of the various components of the SWIPA assessment and the links between them.

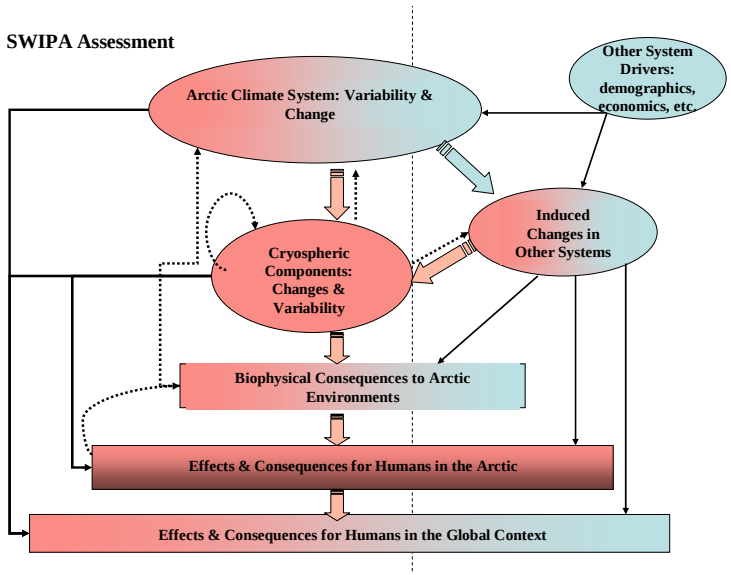

Figure 1.2. The Arctic cryosphere and linkages to various levels of effects or consequences of changes. The SWIPA assessment focuses on the red components highlighted in the various elements of the overall system and the cascade of effects shown as block arrows. Several points emerge from the diagram: (i) Arctic climate change induces change in systems other than just the cryosphere itself (heavy line arrows); (ii) other drivers of the overall system induce changes in various elements (lighter line arrows) (and thus discerning those solely due to cryospheric change may be difficult especially as consequences for humans); (iii) feedbacks are important aspects of the overall system (dotted heavy arrows, and not all are shown), and may induce synergistic as well as antagonistic effects in receiving elements (e.g., human behavioral responses to Arctic cryospheric changes will induce feedbacks in the biophysical system); (iv) climate-induced changes in the cryosphere have differing consequences within versus outside the Arctic; and (v) cumulative effects (i.e., additive or multiplicative cascades through components of the overall system) are implied but not explicitly shown on the figure.

1.4. Roles and relevance of the cryosphere

The cryosphere is a defining aspect of polar and high altitude regions of the globe, with the relative importance of various components differing as to their location on the Earth. The cryosphere is a fundamental regulator or controller of local and regional climate as well as global climate systems. Moreover, the cryosphere itself provides direct services (or impediments) to other elements of the earth system, including humans, as well as indirectly affecting the nature of services provided by those other elements (i.e., ecosystem provisioning and regulatory services).

Both the Arctic generally, and the Arctic cryosphere in particular, combine to act as the thermostat for the Northern Hemisphere. Heat from the warmer lower latitudes is transported to the Arctic by ocean currents and the atmosphere. Air and water masses are cooled in the Arctic and then returned to southern latitudes as cold air outbreaks and cold ocean currents. The Arctic also cools the Earth by reflecting radiation back into space. These processes regulate many aspects of the global climate system. Moreover, general properties of the cold sink enable heat transfer northward (thus cooling more southerly source regions), and with that, physical transport of atmospheric and water constituents (i.e., gases, freshwaters and marine waters). Contaminants, aerosols, dust and soot mostly generated in southerly areas are also transported northward in gases and water, as are nutrients and biota. Nutrients released locally or advected into the Arctic provide the basis for much Arctic productivity and may be concentrated at key frontal regions between southern and northern oceanic water masses. Properties of the cold sink also result in the deposition, precipitation or concentration of these transported constituents, which include anthropogenically produced contaminants. On the longer timescale, storage of greenhouse gases (GHGs), and accumulation of organic material or ground ice has taken place due to natural processes. Gaseous carbon that has been produced in geological deposits is currently capped by permafrost.

The cryospheric components represent the solid phases of water or regular cyclical changes between solid, liquid or gaseous phases. Solid phases provide fundamental physical structuring of the Arctic environment (e.g., ice as a platform for activities) and also act as significant storage reservoirs for many constituents (e.g., water itself as ice or semi-permanent snow, contaminants, impermeable permafrost storing GHGs). Thus, from the perspective of SWIPA, recent cryospheric change primarily represents an irregular shift in phase from solid to liquid, a shift toward greater durations of the liquid phase, or a shift in timing of such phases. Phase changes themselves are key (and essential) structural changes in the physical systems. Many ecosystem components and the resulting services that humans receive from those ecosystems rely upon fixed phases of the cryosphere or upon regular relatively predictable spatio-temporal shifts in such (e.g., seasonally predictable patterns of ice formation in a particular location). Physical services include acting as a stable platform upon which activities (e.g., travel on ice) or infrastructure (e.g., basis for roads) can occur. Alternatively, the liquid phase of water is fundamental to life and affects freshwater supply which, in turn, profoundly affects Arctic ecosystems, their productivity and services, and ecological links to humans, as well as affecting humans directly. More subtle cryospheric services include sources of water to maintain stream flow (e.g., glacial melt), acting as drainage barriers in the landscape (e.g., permafrost) thereby maintaining local water balances, and as elements that re-structure (e.g., river ice jams) or protect (e.g., land-fast sea ice) landscapes. Conversely, depending upon the nature of the human activity, the solid phase of the cryosphere may act as an impediment to some activities or services preventing or altering how or when the environment might be used (e.g., sea ice as a barrier to shipping or resource exploitation). Thus, any observed physical changes in the Arctic cryospheric components represent significant departures from recent norms in the environment from which a cascade of consequences may result.

Box 1.1. Discerning change and variability in complex systems

One generation ago, the concept of climate change was very strict: Most scientists agreed that a statistically significant change from one standard normal period (such as 1961–1990) to another similar period defined by the World Meteorological Organization (WMO) would constitute a change locally.

One decade ago, the criteria for change were relaxed by the Intergovernmental Panel on Climate Change (IPCC, 2001). Owing to new observational evidence, the IPCC stated in its Third Assessment report that "*it is very likely that the 1990s was the warmest decade... in the instrumental record (1861–2000)*".

Less than five years ago, in its Fourth Assessment Report (IPCC, 2007) the IPCC stated: "*The last time the polar regions were significantly warmer than present for an extended period (about 125,000 years ago), reductions in polar ice volume led to 4 to 6 m of sea level rise.*"

In the period since the completion of the ACIA assessment in 2004, the Arctic has experienced its highest temperatures of the instrumental record.

Why has it come this far? First of all it has to do with climate change itself. Changes in the Arctic have taken scientists working in the field by surprise. Neither previous observations nor modeling experiments had indicated that cryospheric change could take place so fast. Second, a number of new observational techniques (such as satellites in the sky, drones in the sea, and automatic stations on glaciers) have led to recent amplification in the number of independent – yet verifiable – pieces of evidence, which all tell the same story: The Arctic cryosphere is changing rapidly. Scientists are now convinced that the emerging picture of a rapidly melting Arctic environment is not a coincidence, but a real and significant change in the climate system of the Earth.

Details about significance may be discussed at length, as has been the case in the scientific literature and in the present report, but it is now becoming very clear that the cryosphere is changing rapidly and that neither observations nor models are able to tell the full story.

The need for concerted monitoring of the cryosphere has emerged with this report as one of the most pressing challenges of our time.

1.5. What SWIPA covers and does not cover

Cryospheric change and variability is fundamentally linked to climate change (see Boxes 1.2 and 1.3) and climatic variability. The SWIPA assessment is not, however, an assessment of climate change *per se*, neither is it a comprehensive update of the ACIA results. Like ACIA, the SWIPA assessment considers the implications of change in the physical components of the Arctic cryosphere on Arctic human populations, and – where relevant – humans living outside the Arctic. The ACIA findings constitute a benchmark for the SWIPA assessment, and an assessment against which the new information presented in the SWIPA assessment can be compared.

Box 1.2. Are Arctic climate and cryospheric changes attributable to anthropogenic causes?

Once change in a system of interest has been observed and confirmed, the second step is to understand the causes for that change. That is, with respect to climate and cryospheric change in the Arctic, is this the result of 'natural' (i.e., non-anthropogenic) drivers and thus part of other system cyclicity or long-term evolution, or is it the result of 'non-natural' (i.e., anthropogenic) drivers?

High association of climate changes with significantly increased levels of anthropogenically produced GHGs, and the understanding of causal linkages between GHGs (a primary driver) and the climate system (a responding system), both provide evidence that human-derived activities have altered, and are continuing to alter the global climate system over the recent past (e.g., see IPCC, 2007, and ACIA, 2005). Accordingly, although the SWIPA Science Report focuses primarily upon Arctic cryospheric changes driven by underlying climate changes rather than upon causation of those changes, the overwhelming basis for conducting such an assessment is that anthropogenic drivers are significant and fundamental contributors to Arctic climate change, which is the main driver of changes in the Arctic cryosphere.

Box 1.3. An example of rapid change in the cryosphere attributed to global warming

The Arctic is warming faster than other regions of the Earth. This is known as the Arctic amplification effect (Figure 1.3). The effect can be local such as through loss of sea ice (Miller et al., 2010) or as a result of the planetary atmospheric and oceanic circulation (Langen and Alexeev, 2007). Forcing of the 20th and 21st century warming is generally attributed to changes in solar heating, volcanism, GHGs, and aerosols. In discussing these impacts it is important go beyond simply independently correlating the time series of the forcing function with the northern hemispheric temperature record, and put the contribution of each factor on a comparative quantitative basis. This was done by Crowley (2000) and the results are summarized in Figure 1.4 showing historical forcing over time, where the forcing influence is normalized in terms of W/m^2. Volcanoes have a cooling influence of up to -5 W/m^2 that can last for a year or two. Increased carbon dioxide (CO_2), a warming influence with a continuing increase in the second half of the 20th century, has a value of 2.4 W/m^2 by 2000. Sulfate (SO_4) is an aerosol with a cooling influence of 0.6 W/m^2 by 2000. Solar forcing has decadal and centennial variability, but its influence over the previous 200 years is below 0.4 W/m^2. In 2000, the ratio of the CO_2 influence to the solar contribution is 8:1. Crowley's work was echoed in the IPCC AR4 Summary for policymakers (IPCC, 2007) which stated that most of the observed increase in global average temperature since the mid-20th century is very likely to be due to the observed increase in anthropogenic GHG concentrations and therefore it is unlikely that the increased warming and melt of ice in the Arctic is only due to the solar component (see Stott et al., 2000 and Overland, 2009 for further discussion).

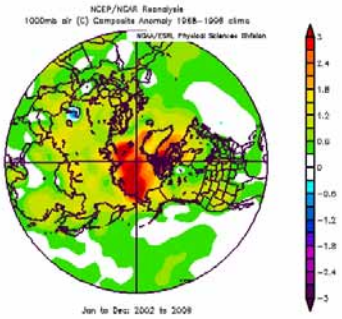

Figure 1.3. Near-surface air temperature anomaly multiyear composites for 2002–2009. Anomalies are relative to the 1968–1996 mean and show a strong Arctic amplification of recent temperature trends. Data are from the NCEP-NCAR Reanalysis through the NOAA/Earth Systems Research Laboratory, generated online at www.cdc.noaa.gov.

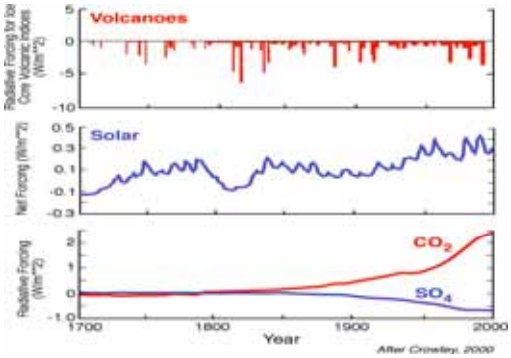

Figure 1.4. Relative forcing of 20th century northern hemispheric temperature increases. Redrawn after Crowley (2000).

The climate change history and future projections presented in the ACIA report (ACIA, 2005) and the IPCC Fourth Assessment report (IPCC, 2007) represent the 'climate framework' for the SWIPA assessment. This information is summarized in Chapter 2, this volume. Development of new climate change projections or detailed analyses of recent trends in anthropogenic GHG emissions were outside the scope of this assessment. New information presented in this assessment concerning, for example, modeling activities is therefore restricted to evaluating models to select those most useful for refining projections for individual cryosphere components.

All parts of the Arctic cryosphere are a product of and subject to climatic conditions, but whereas seasonal phenomena (such as seasonal snow cover and seasonal ice on rivers and lakes) respond to specific weather conditions and rapidly changing climatic conditions, other parts of the cryosphere (such as permafrost) respond very slowly. The implication is that the timescales for which it is meaningful to model change in one part of the cryosphere may be very different from those that are meaningful for modeling change in another. Furthermore, the capability of different models to project change in a given component of the cryosphere may differ substantially. Therefore different model assemblies, scenarios, and timescales have been used in the projections of change in the different components of the Arctic cryosphere.

While reference is made to possible relationships between short-lived climate forcers such as black carbon ('soot') and some components of the Arctic cryosphere, this is not assessed in detail in the present assessment. A separate AMAP report on the state of scientific knowledge concerning short-lived climate forcers of climate change in the Arctic will be presented to the Arctic Council at its meeting in May 2011 (AMAP, 2011).

1.6. The SWIPA assessment process

The SWIPA assessment was produced by more than 200 scientists and experts from the Arctic and non-Arctic countries. They were initially nominated by countries and relevant international bodies and

were selected on the basis of their scientific qualifications by appointed convening lead authors. These experts were charged with compiling and evaluating information from Arctic monitoring networks and recent national and international research activities, such as those carried out during the International Polar Year (2007–2008), focusing on new information gathered since the ACIA assessment. Each chapter of the present assessment was drafted by a group of experts including relevant expertise from different scientific disciplines and geographical areas. A SWIPA assessment 'Integration Team' including the convening lead authors for each chapter was responsible for the overall conduct and organization of the assessment. A strict and independent peer review process was established by the AMAP Working Group to secure and document the integrity of the process.

This assessment report is fully-referenced. The majority of the assessment is based purely on information that is published and available in the peer-reviewed scientific literature or on new results obtained using well-documented models and observational methods. In the case of sections dealing with for instance societal implications of cryospheric change (in the specific chapters and in Chapter 10 in particular) some information sources differ from this principle. In order to assess these aspects, the authors needed to use some 'grey literature' such as government reports, design standards, or anecdotal evidence. However, all materials used in the preparation of this assessment, that are not available in the mainstream published scientific literature, have been collected and are available through the AMAP Secretariat.

The assessments by chapter authors followed recommendations to promote the use of common terminology as far as possible. This included use of terminology associated with probability statements where discussion of future events and conditions need to take into account the likelihood that these conditions or events will occur. To ensure consistency of the summarized material, the procedures used by ACIA (as refined from that of IPCC) were used throughout this report.

Statements regarding the likelihood of particular events or conditions occurring reflect expert evaluation of peer-reviewed results, typically from multiple lines of evidence.

The assessment presented in this report has been subject to a comprehensive review process, which involved a review by national experts that contributed data and information to the assessment, to verify that the interpretation of their data was correct and acceptable to the primary sources. In addition, a full review was conducted by a team of independent international peer-reviewers (see the Preface for further details of the review process).

Documentation of the results of the peer review process applied to the SWIPA assessment is available on the AMAP website: www.amap.no.

1.7. Readers guide: What will the readers find within each chapter?

This report presents the findings of the SWIPA assessment as developed by the report authors, produced under their responsibility. It represents the scientific findings of a large group of independent scientific experts.

A separate report, produced under the responsibility of the AMAP Working Group, summarizes the SWIPA Science Assessment (add reference). A SWIPA Summary for Policymakers produced under the AMAP Working Group and presented to the Arctic Council Ministers at their meeting in Nuuk, Greenland, 2011 includes policy-relevant scientific recommendations (add reference). The scientific

SWIPA assessment report (the present report) provides the validated scientific basis for all statements made in the report and the SWIPA summary for policymakers, as confirmed by the lead authors of the SWIPA scientific assessment.

This report is developed in five main parts. The first part (Chapters 1 to 3) introduces and defines the scope of the assessment, and presents the background climate information that establishes the framework for the assessment of Arctic cryospheric change. The second part (Chapters 4 to 9) describes the physical and other aspects of each of the individual components of the Arctic cryosphere, including discussion of the impacts of change in the cryospheric components concerned. A third part (Chapter 10) focuses on the combined impacts of Arctic cryospheric change on Arctic human society. The fourth part (Chapter 11) focuses on cross-cutting issues of importance at local, regional and global levels. Finally, Chapter 12 presents an integrated synthesis of the findings of the assessment together with the conclusions and recommendations of the assessment as a whole.

A more detailed description of the content of each of the main chapters is as follows:

- Chapter 1 sets the stage for the assessment describing its scope and how it was accomplished.
- Chapter 2 presents an overview of past and present climate in the Arctic, establishing the 'climate framework' for the assessment.
- Chapter 3 provides a critical evaluation of the climate models employed by the IPCC in relation to their ability to produce projections for components of the Arctic cryosphere.
- Chapter 4 describes the snow component of the cryosphere, focusing on variation and change in snow extent and depth over time, seasons and geographical extent.
- Chapter 5 presents a status report for permafrost conditions and the consequences of the permafrost thawing that is ongoing in different regions of the Arctic.
- Chapter 6 describes changes in Arctic river and lake ice and how these affect biological systems and the human use of these components of the cryosphere.
- Chapter 7 describes changing Arctic mountain glaciers and ice caps, the changes that are being observed and the mechanisms behind these changes.
- Chapter 8 presents a review of the status of the mass balance of the Greenland Ice Sheet, updating a preliminary assessment of this subject that was prepared for the UN Climate Change Conference 2009 (UNFCCC COP15) in Copenhagen (AMAP, 2009).
- Chapter 9 presents a synthesis of observed changes in Arctic sea-ice extent and thickness that have been recorded over the past several decades. The chapter also assesses how the observed changes affect biological life associated with sea ice and effects on human activities, such as shipping and traditional hunting.
- Chapter 10 presents a synthesis of available information on the potential effects of the changes in the Arctic cryosphere on humans and human activities.
- Chapter 11 presents an integrated assessment of cross-cutting scientific issues that affect multiple components of the cryosphere. These include assessments of feedback mechanisms leading to global sea-level rise; ways in which cryospheric change affects transport and bioavailability of contaminants within the Arctic region; ways in which cryospheric change affects Arctic ecosystems; and systems available for observing changes in the Arctic cryosphere.
- Chapter 12 presents an overall summary of the major findings of the SWIPA assessment. It is based on the logical consequences of and conclusions stemming from the scientific findings presented in the preceding chapters.

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2. Arctic Climate: Recent Variations

John E. Walsh, James E. Overland, Pavel Y. Groisman, Bruno Rudolf

Contents

Key findings

- Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided. Still to be decided.
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Summary

In the period since the completion of the Arctic Climate Impact Assessment in 2004 (ACIA, 2005), the Arctic has experienced its highest temperatures of the instrumental record, even exceeding the warmth of the 1930s and 1940s. Recent paleo-reconstructions, while subject to uncertainties in the methodology and spatial representativeness, also show that Arctic summer temperatures have been higher in the past few decades than at any time in the past 2000 years. The geographical distribution of the recent warming points strongly to an influence of sea ice reduction on temperature, as the greatest temperature increase has occurred in the lower atmosphere over the marginal sea ice zone during autumn. The seasonality and location of the maximum warming indicates that the Arctic may have passed the threshold at which absorption of solar radiation during summer limits ice growth during the following autumn and winter, initiating a feedback that leads to a substantial increase in Arctic Ocean surface air temperatures. A secondary maximum of warming during springtime is consistent with an earlier loss of terrestrial snow cover in recent years. The spatial pattern of the near-surface warming shows the signature of the Pacific Decadal Oscillation (PDO) in the Pacific sector as well as the influence of a dipole-like circulation pattern in the Atlantic sector. Since the Arctic Oscillation (AO) has been generally neutral, it has not been a major determinant of Arctic warmth of the past several years. Spatially averaged Arctic precipitation over the land areas north of 55° N shows large year-to-year variability. This variability is superimposed on an increase of about 5% since 1950. While the statistical significance of the precipitation increase has not been demonstrated, the years since 2000, and especially since 2004, have been quite wet according to both precipitation and river discharge data. The five wettest years since 1950 have all occurred in the past decade. Estimates of precipitation over the Arctic Ocean remain a high priority need in assessments of recent Arctic change. There are indications of increases in cloudiness over the Arctic, especially in low clouds during the warm season, consistent with a longer summer season and a reduction of summer sea ice. However, conclusions about trends in Arctic cloudiness are limited by large uncertainties in Arctic cloud data.

While there are indications of local changes in the occurrence of storm events and extreme temperatures in Alaska, systematic changes in strong storms and other extreme events have not been documented on a pan-Arctic basis. In view of the potential impacts of extreme events on the cryosphere as well as on human activities, research on changes in climate-related extreme events in the Arctic must be considered a priority.

The Arctic Ocean has experienced enhanced oceanic heat inflows from both the North Atlantic and the North Pacific. The Pacific inflows appear to have played a role in the retreat of sea ice in the Pacific sector of the Arctic Ocean. The association between sea ice and enhanced North Atlantic layer heat content is complicated by the strong halocline above the Atlantic layer, so the role of the Atlantic water heat anomalies is still under investigation. Nevertheless, the Atlantic water heat influx to the Arctic Ocean appears to be characterized by increasingly warm pulses separated by brief respites of cooling, as occurred in 2008/2009.

2.1. Introduction

The cryospheric changes described in this assessment report are part of a broader suite of interrelated variations in the Arctic climate system. A thorough review of Arctic climate, its relation to external forcing, and its variations through 2004 were presented in the Arctic Climate Impact Assessment report (ACIA, 2005: chapter 2). As a result, this report includes a brief overview of more recent Arctic climate variations, by describing Arctic climate variations during the post-ACIA period and by summarizing key observational studies of Arctic climate that have appeared since 2004. This chapter emphasizes the primary climate drivers (temperature, precipitation, storminess, clouds, ocean) of cryospheric change. Recent changes in the primary cryospheric variables (snow, sea ice, glaciers, permafrost) are presented in subsequent chapters. This chapter does not project recent variations into the future, as the main sources of such projections are the climate models discussed in Chapter 3.

2.2. Air temperatures

- In the period since the completion of the Arctic Climate Impact Assessment in 2004, the Arctic has experienced its highest surface air temperatures of the instrumental record, exceeding even the warmth of the 1930s and 1940s.
- Subject to the uncertainties inherent in proxy information, recent paleo-reconstructions show that Arctic summer temperatures have been higher in the past few decades than at any time in the past 2000 years.
- The spatial distribution of the recent warming points strongly to sea-ice reduction influencing warming, as the greatest temperature increase has occurred in the lower atmosphere over the marginal sea ice zone during autumn.
- A secondary maximum of warming during springtime is consistent with an earlier loss of terrestrial snow cover in recent years.
- The spatial pattern of the near-surface warming shows the signature of the Pacific Decadal Oscillation in the Pacific sector as well as the influence of a dipole-like circulation pattern in the Atlantic sector.

The Arctic Climate Impact Assessment (ACIA) reviewed Arctic climatic history from the pre-Quaternary through the Holocene (ACIA, 2005: section 2.7). Since the publication of the ACIA report, Arctic paleo-climate studies have refined temporal and regional variations of Arctic paleo-climate. A recent integrative study provided a reconstruction of pan-Arctic summer temperatures over the past 2000 years based on various proxies, including lake sediments, pollen records, diatoms, and tree rings (Kaufmann et al., 2009). This reconstruction adds to the previous knowledge base by showing that the Arctic had been undergoing a slow (summer) cooling for most of the 2000-year period prior to the 1800s (Figure 2.1). This cooling is consistent with the slow variations of the Earth-Sun orbital parameters, which affect the solar radiation reaching the Arctic in the sunlit portion of the year. However, warming since the 1800s, as shown by the instrumental data (Figure 2.1), has left the Arctic warmer by a considerable margin than at any time in the preceding 2000 years. The recent instrumental temperatures are outside the envelope of the natural variability depicted by the reconstruction. For example, the warming is far more than simply a recovery from the so-called Little Ice Age, which is apparent from the 1500s through the 1800s (Figure 2.1). While the figure conveys a visually striking picture of recent Arctic warming, it should be noted that the data used by Kaufman et al. (2009) were biased towards Greenland, with very few sites in Siberia and no data from the polar ocean. Moreover, when uncertainties from the methods and the sparseness of the data are included, the recent decades of proxy data (Figure 2.1) are not significantly warmer than 2000 years ago. Finally, the locations of the proxy sites generally do not correspond with the instrumental data locations, thus contributing to the differences between the proxy and instrumental values in the mid-20th century.

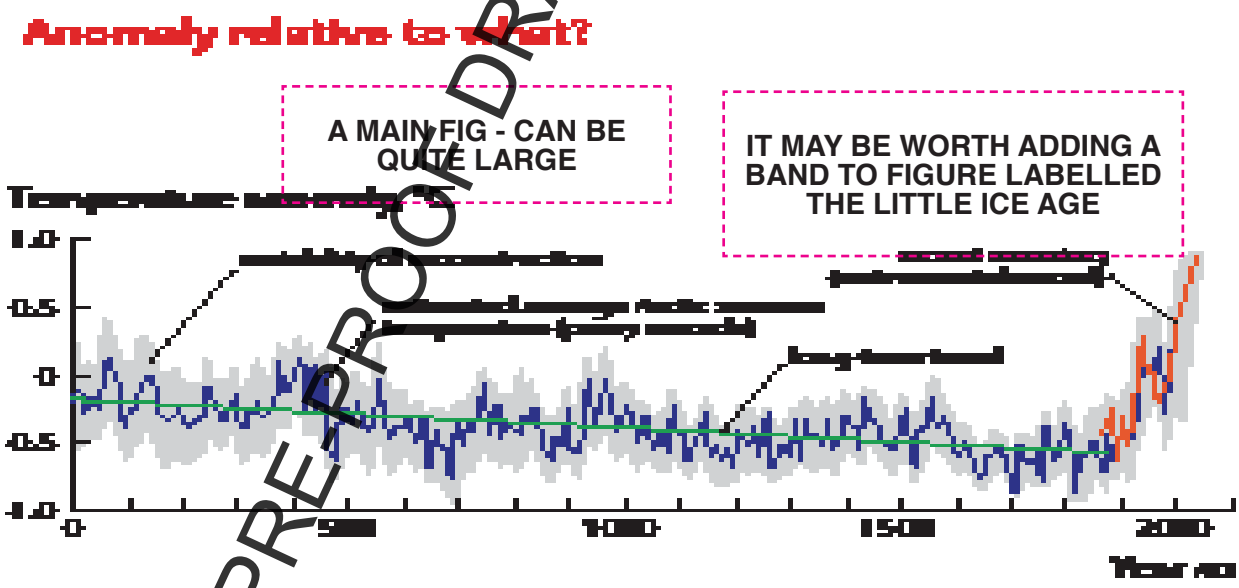


Figure 2.1. Estimated Arctic average summer air temperature anomalies for the past 2000 years (relative to the 1961 to 1990 mean), based on proxy records from lake sediments, ice cores, and tree rings. The shaded area represents variability among the 23 sites used for the reconstruction. Source: Kaufman et al. (2009), modified by UCAR (University for Atmospheric Research).

The recent warming is highlighted in greater temporal detail in Figure 2.2, which shows the annual temperatures averaged over a pan-Arctic domain (60° to 90° N) up to 2009. Consistent with the long-term reconstruction (see Figure 2.1) and with the Arctic temperature depictions presented in the ACIA report (ACIA, 2005), Figure 2.2 clearly shows the early 20th century warming, followed by the mid-century cooling and the late century warming. It is notable, however, that the warmest five years in the entire record have all occurred since 2005, i.e. in the post-ACIA period. Monte Carlo tests (1000 trials in which the data points in Figure 2.2 are randomly reordered) show that the likelihood of such a sequence occurring by random chance is close to zero. Including the past five years of temperature data in the temperature record takes the recent Arctic warming from a state of temperatures comparable to those of the 1930s to a state that is warmer than the 1930s. This is one of the more notable occurrences in Arctic climate for the post-ACIA period and is consistent with the reductions in sea ice, snow cover, and glacier mass discussed in Chapters 4, 5, and 7. Moreover, because the time series is based on land station observations only, it does not explicitly use temperature measurements from the ocean areas where sea ice has been lost over the past several years. The recent anomalies (see Figure 2.2) may therefore be conservative estimates of the pan-Arctic values during the seasons (summer and autumn) of greatest sea ice loss.

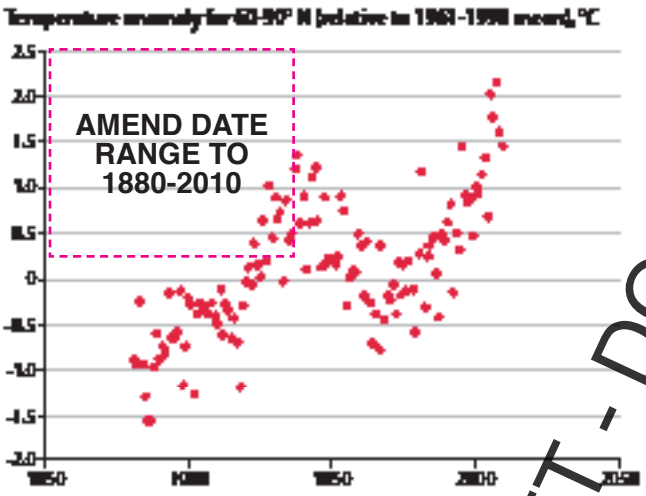


Figure 2.2. Annual Arctic surface air temperature anomalies for 1880 to 2009 (relative to the 1961 to 1990 mean), averaged over the area between 60° and 90° N. Source: P. Groisman, NOAA/National Climatic Data Center.

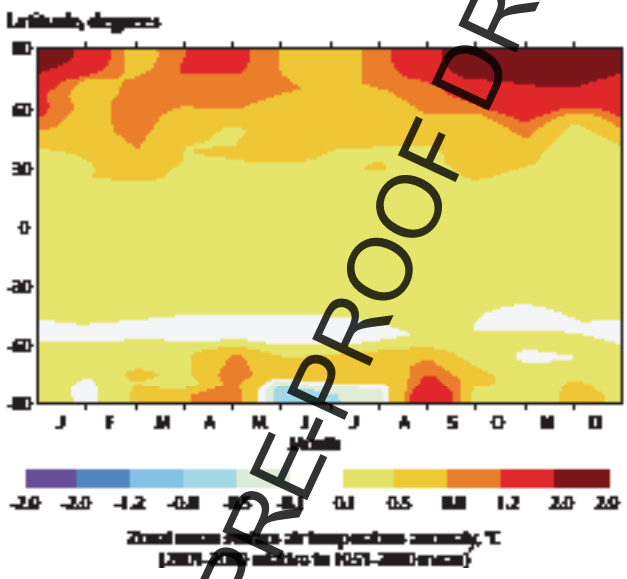


Figure 2.3. Monthly surface air temperature anomalies averaged over the period 2001 to 2009 (relative to the mean for 1951 to 2000) shown as a function of latitude. Source: NASA Goddard Institute for Space Studies (http://data.giss.nasa.gov/gistemp).

The recent Arctic warming varies with season and is stronger than the warming at mid- and lower latitudes (Figure 2.3). It is also stronger than the warming over Antarctica in all calendar months except August and September, when the warming near both poles is comparable. Of particular note is the seasonality of the Arctic warming, which is greatest over the Arctic Ocean during autumn and early winter. This seasonality is consistent with the recent loss of Arctic sea ice, and serves as an indication that the ice-albedo feedback has emerged as a contributor to temperature anomalies in the Arctic. A secondary maximum of Arctic warming propagating poleward from 50° to 60° N during spring is consistent with an earlier seasonal loss of snow cover over northern land areas (see also Section 5.3.1.2). The seasonal variation of the warming is consistent with greenhouse-driven changes projected by global climate models (see Chapter 3).

The role of the ice-albedo feedback in the Arctic temperature record of the past five years is further supported by the spatial patterns in the annual (Figure 2.4) and seasonal (Figure 2.5) temperature anomalies of the past five years relative to the mean temperatures for 1951 to 2000. The annual pattern contains anomalies exceeding 2 °C over much of the Arctic Ocean, and shows a clear polar amplification (Figure 2.4). It should be noted that reduced sea-ice concentrations and thicknesses, in addition to reduced sea-ice extent, can contribute to the feedback between sea ice and surface air temperature over the central Arctic Ocean. Equatorward of the Arctic Ocean, the warming is generally stronger over the continents than over the oceans. These spatial features are consistent with greenhouse-driven projections of change simulated by climate models (Chapter 3).

The seasonal air temperature patterns (Figure 2.5) highlight the maximum warming in autumn and winter, expanding on the patterns of warming by latitude and season shown in Figure 2.3. The seasonal patterns, especially for winter, contain more spatial variability than the annual pattern, consistent with advective influences arising from anomalies of the atmospheric circulation. The seasonal patterns in Figure 2.5 show two differences from the corresponding patterns depicted in the ACIA report (ACIA, 2005): a maximum warming over the marginal ice zone during autumn and winter, and some warming over the Arctic Ocean during summer – despite the large thermal capacity of the ocean. The summertime warming of the Arctic Ocean is consistent with thinner sea ice and/or an earlier retreat of sea ice during summer (see Chapter 9).

To some extent, these recent spatial patterns in temperature change (Figure 2.5) are shaped by the phase of the low-frequency (decadal or multi-decadal) variations in atmospheric circulation. Two large-scale modes for which there are documented effects on regional Arctic air temperatures are the AO, which drives temperature anomalies from eastern Canada across the North Atlantic to northern Eurasia (Thompson and Wallace, 2000), and the PDO, which has a strong influence on sub-Arctic temperatures in the Pacific sector (Mantua and

Hare, 2002). The Arctic warming of the late 1980s and early 1990s has been attributed by various researchers (e.g., Comiso, 2003; Overland et al., 2008) to a predominantly positive phase of the AO (Figure 2.6). And in fact, the warming of the 1980s to 1990s was indeed stronger over northern Eurasia than over many other sectors of the Arctic. In contrast, the more recent Arctic warming cannot be attributed to the AO. First, the AO index has been in a generally neutral state (oscillating between positive and negative phases) since 1997 – despite the Arctic's warmest years in the instrumental record having occurred since 2004 (see Figure 2.2). Second, the AO index reached the most negative values ever recorded in December 2009 to January 2010. At the same time, the High Arctic was relatively warm while northern Europe and Asia experienced a period

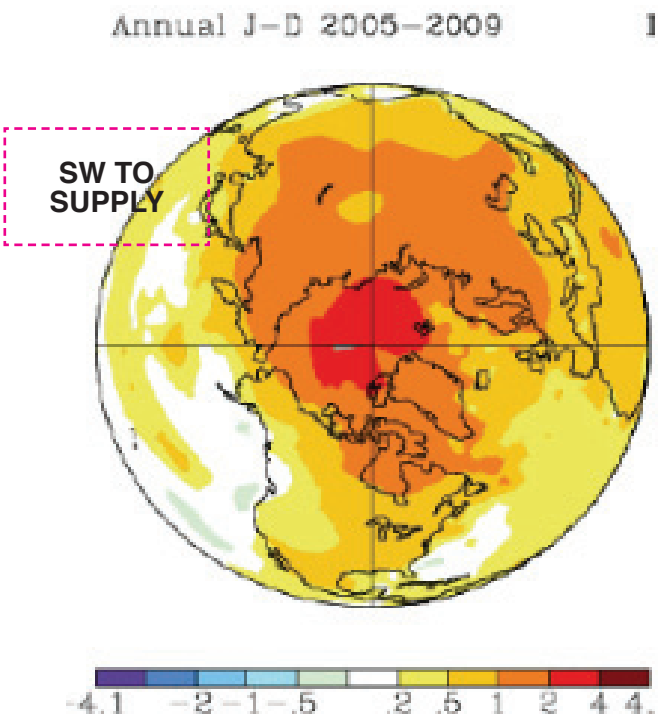


Figure 2.4. Annual surface air temperature anomalies averaged over the period 2005 to 2009 (relative to the mean for 1951 to 2000). Source: NASA Goddard Institute for Space Studies (http://data.giss.nasa.gov/gistemp).

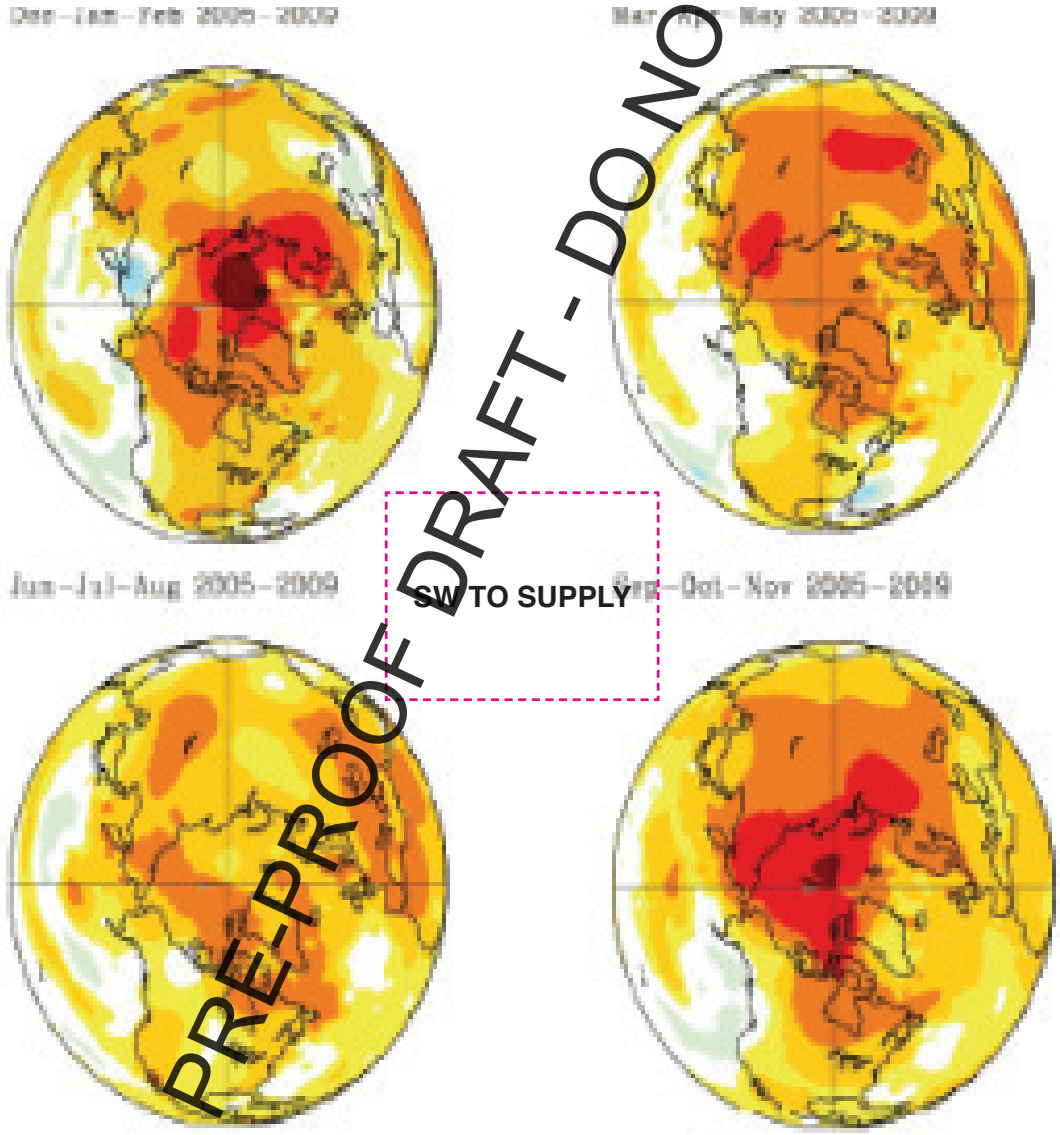


Figure 2.5. Seasonal surface air temperature anomalies averaged over the period 2005 to 2009 (relative to the mean for 1951 to 2000): winter, upper left; spring, upper right; summer, lower left; autumn, lower right. Source: NASA Goddard Institute for Space Studies (http://data.giss.nasa.gov/gistemp).

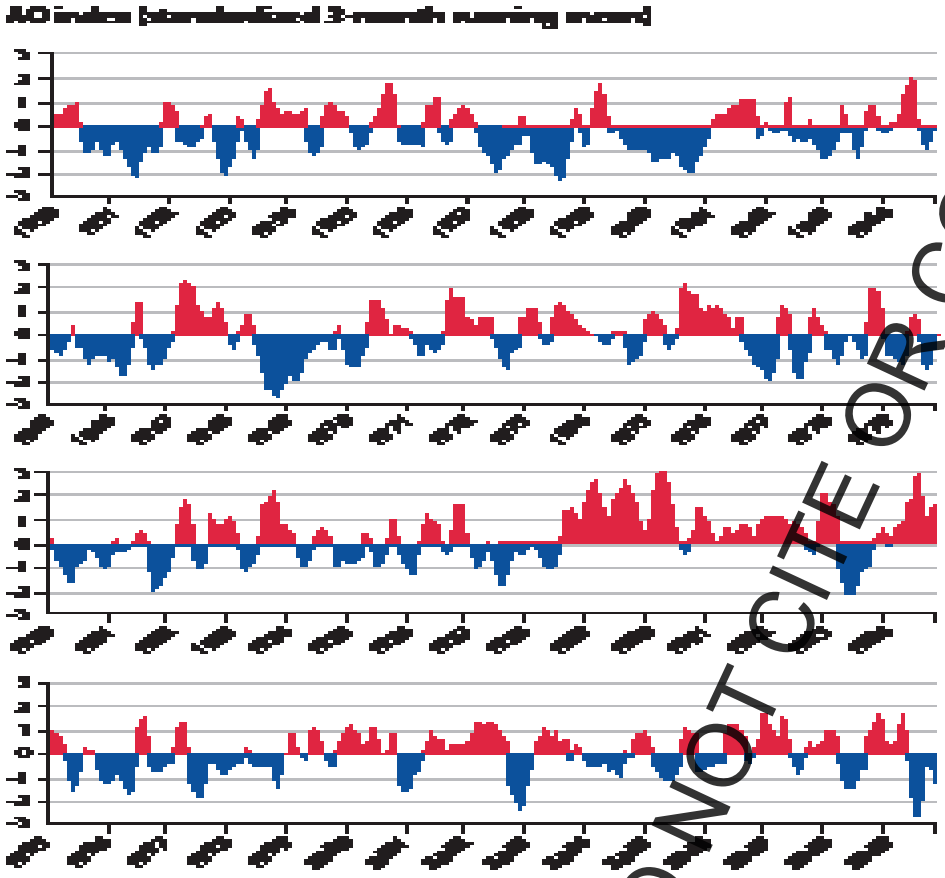


Figure 2.6. Arctic Oscillation index, plotted as a standardized 30-month running mean for the period 1950 to 2009. Source: NOAA/Climate Prediction Center.

of extreme cold. These concurrent anomalies of opposite sign point to the perils of using spatially aggregated temperature data as proxies for temperatures in particular sub-regions. For example, in the context of the snow, water, ice, and permafrost in the Arctic, a seasonal or annual temperature anomaly averaged over the Arctic may differ even in sign from a local or regional anomaly most relevant to a particular glacier, ice sheet, lake, river basin, or sub-Arctic sea.

The PDO has been shown to be a primary determinant of wintertime temperature anomalies in northwestern North America (Hartmann and Wendler, 2005) and it is clear from Figure 2.7 that the PDO index (evaluated from Pacific sea-surface temperatures) does indeed exhibit multi-decadal variability. The increase from the mid-1970s to the early 1980s corresponds with a substantial increase in air temperature over Alaska and northwestern Canada. Whereas, the negative values of 2008 to 2009 coincide with an episode of below-normal air temperature in 2008 to 2009, although the PDO index became positive when averaged over the final six months of 2009. The influence of the PDO extends westward to far eastern Siberia, where temperature anomalies are out of phase with those of Alaska and the Yukon, largely as a result of the intensification cycles of the Aleutian Low pressure system in conjunction with the PDO. Indeed, the coupler of temperature anomalies of opposite sign in the winter pattern (see Figure 2.5) and in the multi-decadal trends (ACIA, 2005: section 2.6.2.1) are driven, to a large extent, by the PDO and associated wind anomalies in the vicinity of the Aleutian Low. A major challenge in anticipating temperature changes in the Pacific sub-Arctic is related to the inability to predict phase transitions of the PDO. However, it is notable (see concluding discussion in Section 2.7) that neither the PDO nor the AO has been in a phase conducive to Arctic warming during the past several years – despite the anomalous pan-Arctic warmth of these years (Figure 2.2).

The relatively high Arctic air temperatures of recent years have also been associated with atmospheric circulation patterns conducive to the export of older, thicker sea ice from the Arctic Ocean to the North Atlantic. Several post-ACIA studies have pointed to the prominent role of similar circulation patterns, which have been assigned names ranging from the 'Dipole Anomaly' (Wu et al., 2006) to the 'Arctic Rapid-Change pattern' (Zhang et al., 2008). These patterns are best developed in the winter half of the year but can affect sea-ice export in all seasons. In addition, they are argued to have preconditioned the Arctic sea-ice cover for the rapid summer retreat of the late 2000s (Smedsrud et al., 2008). Overland and Wang (2005) and Overland et al. (2008) highlighted the meridional (across-pole) character of this atmospheric pattern, which in addition to affecting sea-ice export, advects heat into the Arctic Ocean in a pattern distinct from the PDO and AO. More recently, Overland and Wang (2010) have presented evidence that the loss of sea ice has become sufficient to influence the atmospheric heat budget and circulation pattern in the autumn and early winter months. Francis et al. (2009) arrived at a similar conclusion based on a data analysis encompassing a larger sample of years.

A major topic of attention in the past few years has been the vertical structure of the recent Arctic warming, since the vertical structure provides clues as to the nature (drivers) of the warming. Graversen et al. (2008) argued that an elevated maximum of the warming precludes a major role of surface heating, although several subsequent studies (Bitz and Fu, 2008; Grant et al., 2008) have provided evidence of a surface-based warming. The apparent discrepancy between the different analyses is due to the use of different datasets (re-analyses)

and time periods. While the warming is clearly strongest at the surface during autumn in the NCEP/NCAR re-analysis (Kalnay et al., 1996) (Figure 2.8), this near-surface warming is not apparent in the European ERA-40 database used by Graversen et al. (2008). Moreover, the data on which Figure 2.8 is based include the years of extreme ice minima that were not in the Graversen and co-workers study, giving credence to the argument that the ice-albedo feedback to Arctic temperatures is just now emerging in the post-ACIA period (Serreze et al., 2008).

Finally, the studies of variation and trends in Arctic temperature have focused almost exclusively on monthly, seasonal, or annual mean temperatures. There has been little work on systematic changes in variability or extremes. Among the few studies of this kind, Walsh et al. (2005) found little evidence of increased variance in daily temperatures in Alaska and western Canada over the 50-year period ending in 2000. However, there were indications of an increased frequency of daily extreme temperatures from the 1950s to the 1990s. Whether this trend has continued into the past five years of record Arctic warmth (Figure 2.2) is unknown, but model projections indicate that increasing frequencies of record-high daily temperatures are characteristic of climate change driven by increasing greenhouse gas concentrations (Timlin and Walsh, 2007). Given the impacts of extreme events on humans, ecosystems, and other parts of the cryosphere, a priority for research is to determine the relationship between changes in means and extremes of Arctic climate variables.

2.3. Precipitation

- Although the trends are not statistically significant because interannual variability is large, recent annual pan-Arctic precipitation generally exceeds the mean of the 1950s by about 5%.
- The years since 2000 have been quite wet in the Arctic according to both precipitation and river discharge data. The five wettest years since 1950 have all occurred in the past decade.

Evidence for trends in Arctic precipitation is complicated by inadequacies both in *in situ* measurements and remote sensing-derived estimates of precipitation in cold climates. While these deficiencies were highlighted in the ACIA report (ACIA, 2005) and are discussed in Chapter 11, Section 11.6 (Key et al.), it is necessary to reiterate the challenges created by changing station distributions and gauge undercatch of snow. Moreover, high-latitude precipitation gauges are sited preferentially in low-elevation areas. These factors impede attempts to construct temporally homogeneous records of spatially averaged precipitation. Partly for this reason, variations in Arctic precipitation have been examined using atmospheric re-analysis output, either as directly simulated by models (Serreze et al., 2005) or as moisture flux convergences (e.g., Peterson et al., 2006).

The most comprehensive collection of *in situ* precipitation data belongs to the Global Precipitation Climatology Center, which has produced Arctic precipitation time series such as that shown in Figure 2.9 for monthly precipitation averaged over the land areas north of 55° N. The data show a strong

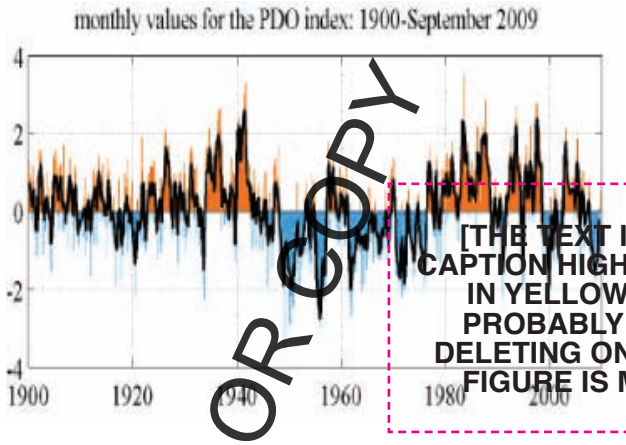


Figure 2.7. Pacific Decadal Oscillation index, expressed in terms of monthly values (thin line) and a five-month running mean (thick line and red/blue shading). Source: Joint Institute for the Study of the Atmosphere and Ocean, NOAA/University of Washington (http://jisao.washington.edu/pdo).

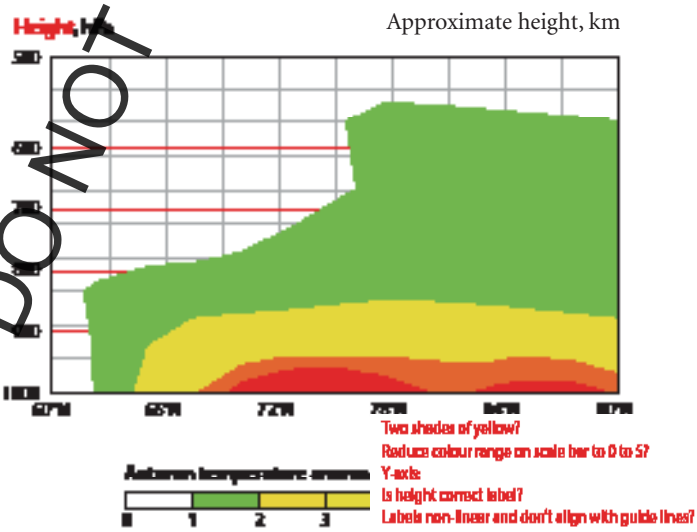


Figure 2.8. Latitude-height cross-section of pan-Arctic autumn (October / November / December) temperature anomalies for 2003-2008 (relative to the 1858-96 base period). Source: J. Overland, NOAA; based on the National Centers for Environmental Prediction / National Center for Atmospheric Research (NCAR/NCEP) re-analysis.

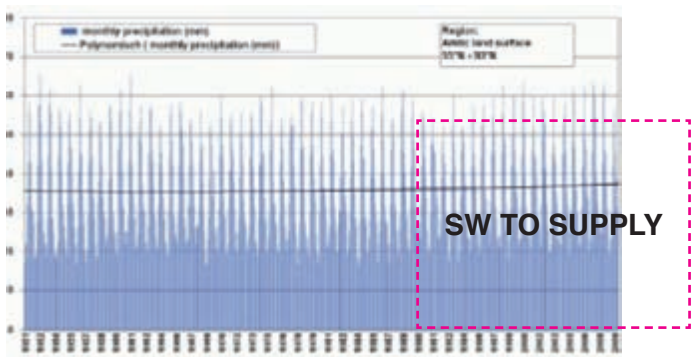


Figure 2.9. Monthly precipitation averaged over land areas north of 55° N for the period ??? to ???. The solid black line shows a least-squares polynomial fit. Source: Global Precipitation Climatology Center / World Meteorological Organization / Deutscher Wetterdienst (Courtesy of B. Rudolf).

seasonal cycle (with larger amounts in the warm season) and considerable interannual variability. A polynomial fit to the data indicates a small increase of about 5% (from 35 mm to 37 mm) over the period 1951 to 2009. This is a modest increase in relation to the variability and is not statistically significant, pointing to the difficulty of extracting significant signals from highly variable precipitation data.

In terms of annual variations in precipitation over the period 1951 to 2009, the five wettest years (exceeding 450 mm) have all occurred in the most recent decade: 2000, 2002, 2005, 2007, and 2008 (Figure 2.10). Anomalies in both hydrological winter and hydrological summer have contributed to these large precipitation amounts in recent years. As noted in Chapter 3, global climate models project an increase in Arctic precipitation under all scenarios of greenhouse forcing. However, in the models as in the observational data, interannual variability is large, resulting in a smaller signal-to-noise ratio for precipitation than for surface air temperature.

The seasonal evolution of the annual precipitation anomalies over the past two decades illustrates the general wetness of the Arctic during the post-ACIA period (Figure 2.11). The summer precipitation anomalies contributed most to the cumulative annual anomalies in 2005 and 2007, although winter and spring provided the larger contributions in 2008. As previously noted, these years were not distinguished by large amplitudes of the AO index or PDO index, although the Arctic Rapid-Change Pattern was prominent from 2006 onward (Zhang et al., 2008). While the predominance of positive anomalies in precipitation amount is apparent, there are occasional years with small overall deficits in the past decade (e.g., 1999, 2002). The cumulative anomalies for most years retain the same sign through the full year, and the years with the largest positive anomalies tend to accumulate the greatest anomaly amounts during summer and autumn.

Finally, the increasing frequency of wet years at high northern latitudes is supported by increases in river discharge amounts (Figure 2.12). For Eurasia, the discharge of the largest rivers has increased by about 10% since 1935, despite the large

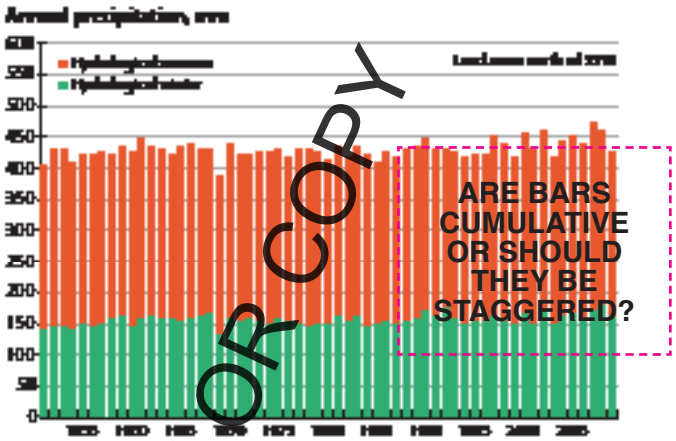


Figure 2.10. Annual precipitation for the period 1951 to 2009 averaged over land areas north of 55° N for hydrological winter (October to March) and hydrological summer (April to September). Source: Global Precipitation Climatology Center / World Meteorological Organization / Deutscher Wetterdienst (Courtesy B. Rudolf).

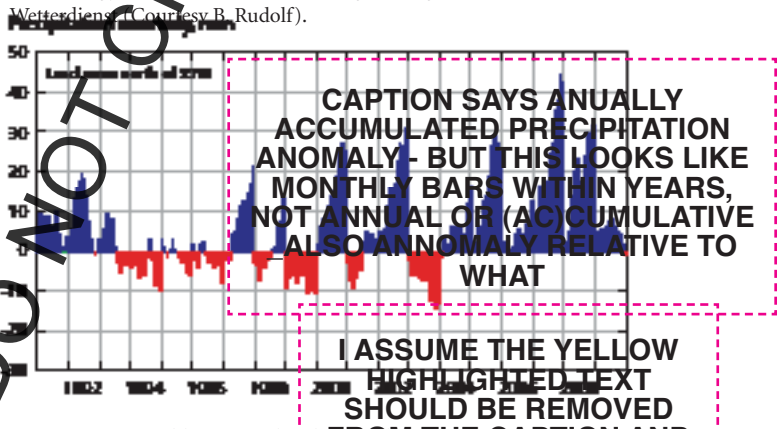


Figure 2.11. Monthly-accumulated anomalies of precipitation for the period 1991 to 2009 (relative to the corresponding monthly means for 1951 to 2000) averaged over land areas north of 55° N. Blue indicates precipitation surplus, red indicates precipitation deficit. Source: Global Precipitation Climatology Center / World Meteorological Organization / Deutscher Wetterdienst (Courtesy of B. Rudolf).

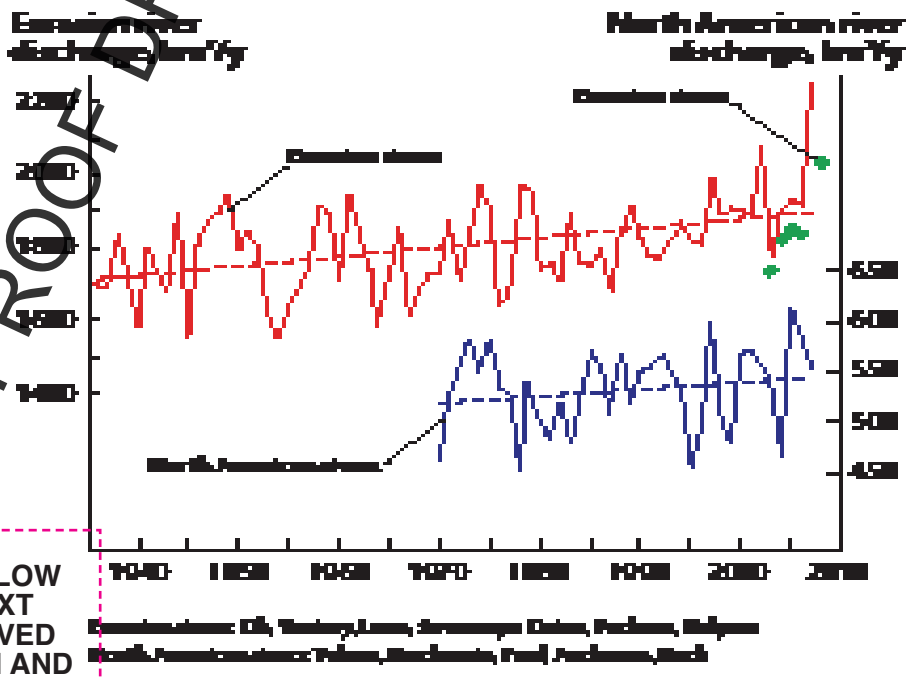


Figure 2.12. Total annual river discharge to the Arctic Ocean from the six largest rivers in the Eurasian Arctic for the observational period 1936 to 2007 (updated from Peterson et al., 2002) and from the five large North American pan-Arctic rivers for the period 1973 to 2006. The least squares linear trend lines are shown as dashed lines. Provisional estimates of annual discharge for the six major Eurasian Arctic rivers are based on near real time data (http://RIMS.unh.edu) are shown as green dots. Source: Richter-Menge and Overland (2009). Courtesy of A. Shiklomanov.

interannual variations apparent. The rate of increase for North America is similar, although the record length of river discharge is shorter. The discharge curves for the two continents show a positive correlation, and their extreme years also show some correspondence with the annual precipitation amounts (see Figures 2.9 and 2.10). Since Arctic river discharge has direct ties to many of the cryospheric variables, it will be discussed further in subsequent chapters of this report.

2.4. Storminess

- Storm activity and the occurrence of temperature extremes have increased at some locations in the North American Arctic, but there are no indications of systematic increases in storminess in the Arctic over the past half century.

In addition to their highly publicized impacts on coastal regions and coastal residents, storms affect the cryosphere through their associated precipitation (affecting glaciers, ice sheets, snow cover, and even permafrost), winds (affecting sea ice motion and the distribution of snow on land and sea ice), and waves (affecting coastal permafrost). While storms have received increased diagnostic analyses through case studies (e.g., Roberts et al., 2008), there have been few rigorous evaluations of variations and trends in storminess in the Arctic, particularly the central Arctic. Wang et al. (2006) reported a northward shift of cyclone activity, primarily during winter, over Canada during 1953 to 2002, while Mesquita et al. (2010) found that temporal trends of cyclones in the North Pacific have generally been weak over the 60-year period ending in 2008, although the U.S. Global Change Research Program (Karl et al., 2009) points to an increase in storminess on the northern Alaskan coast and to associated risks of flooding and coastal erosion. Since any increases in coastal flooding and erosion are also related to retreating sea ice, the role of storminess in itself can be difficult to unravel. Nevertheless, it is apparent from the absence of a comprehensive (pan-Arctic) evaluation of recent variations in storminess that there is a need for systematic assessments of storminess in the Arctic. Such an assessment should include historical variations and their diagnosis, as well as more substantive attempts to project changes into the future. The simulation of future changes in storminess is one of the major challenges facing coarse-resolution global models that are used in assessments such as those of the Intergovernmental Panel on Climate Change (Solomon et al., 2007).

2.5. Cloudiness

- While cloud data for the Arctic are difficult to interpret quantitatively, there are indications of increases in cloudiness over the Arctic, especially in low clouds during the warm season.

Through their large contributions to the surface energy budget, Arctic clouds can have important impacts on the surface energy budget and the cryosphere. These impacts can be manifest in interannual variations as well as trends. For example, Kay et al. (2008) showed that the extreme retreat of sea ice in the summer of 2007 was accompanied by unusually clear skies over much of the Arctic Ocean. Trends and other longer-term variations have been addressed in several post-ACIA studies, although the observational challenges posed by Arctic clouds must be recognized, both for remote sensing and for *in situ* measurements.

Wang and Key (2005) used high-resolution infrared (Advanced Very High Resolution Radiometer, AVHRR) satellite imagery to compute trends of -6% for winter, +3% for spring, +2% for summer, and -2% for autumn per decade during 1982 to 1999. Eastman and Warren (2010), on the other hand, used surface-based observations from 1991 to 2007 and obtained small positive trends in all seasons. Low clouds were primarily responsible for these trends. Perhaps more importantly for cryospheric changes, clouds over sea ice showed a tendency to increase with increasing air temperature and decreasing sea ice in all seasons except summer. Particularly in autumn, there was an increase in low clouds consistent with reduced sea ice, indicating that recent cloud changes may be enhancing the warming of the Arctic and accelerating the decline of sea ice (Eastman and Warren, 2010). This suggestion is consistent with the recent model-based results of Vavrus et al. (2010), who found that in ensembles of 21st century projections by the Community Climate System Model (CCSM3), clouds increased in autumn and decreased in summer during periods of rapid sea ice loss. This seasonality of the sea ice / cloud associations is not inconsistent with the loss of sea ice in recent years such as 2007, and could amplify the loss of sea ice in the future.

2.6. Ocean variations

- The Arctic Ocean has experienced enhanced oceanic heat inflows from both the North Atlantic and the North Pacific. The Pacific inflows appear to have played a role in the retreat of sea ice in the Pacific sector of the Arctic Ocean.
- North Atlantic inflows to the Arctic Ocean appear to be characterized by increasingly warm pulses of water. One such pulse moved along the Siberian shelf break in the mid-2000s, while the next pulse is now poised to enter the Arctic Ocean through Fram Strait.

A key driver of cryospheric change is the variability of the high-latitude oceans. For example, the heat content of the polar oceans directly affects sea ice, tidewater glaciers and ice shelves, snowfall over the high latitudes, and perhaps even the large-scale atmospheric circulation. Salinity variations affect the stratification and control the locations of deep mixing of the oceans, while high-latitude ocean currents contribute to the driving of sea-ice motion and the advection of heat and freshwater anomalies. Historically, the high-latitude oceans have been woefully undersampled by observations, especially below the surface. However, during the past decade, and especially in the post-ACIA period of the International Polar Year (IPY), 2007/2008, there have been unprecedented opportunities to monitor the Arctic Ocean and its exchanges with the mid-latitudes – precisely during a period of unprecedented change in various components of the cryosphere.

Insights into high-latitude ocean variability, as gleaned from preliminary analyses of IPY-period observational data, have been summarized by Dickson and Farbach (2010). Following are brief summaries of those insights that have potential implications for cryospheric variations documented elsewhere in this report. First, the annual mean northward transport

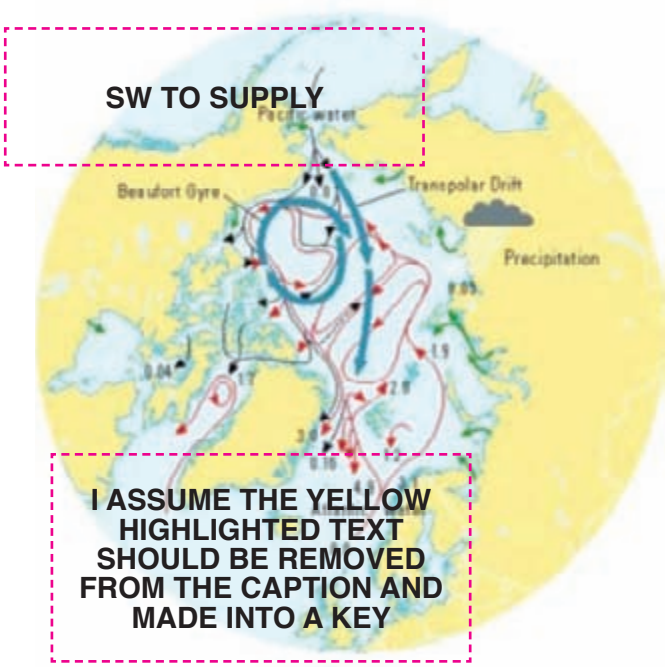


Figure 2.13. Major surface ocean currents and water transport in the Arctic. Black numbers indicate water transport in Sverdrups (106 km3 per second). Mean surface currents are shown by blue arrows, Atlantic water and intermediate currents by red arrows, and Pacific water water by black arrows. Figure courtesy of the Arctic Monitoring and Assessment Programme.

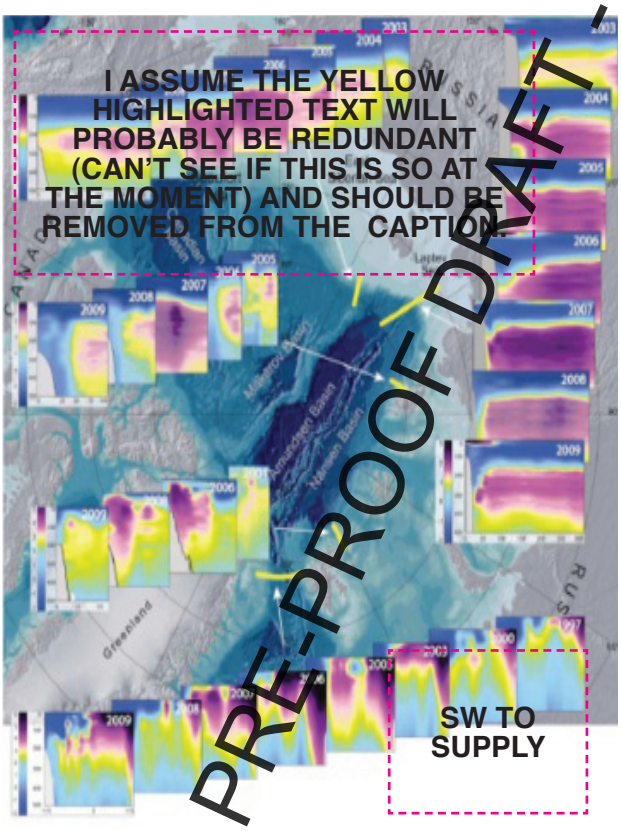


Figure 2.14. Cross-sections of temperature in the upper layers of the Arctic Ocean. Depths are in metres, horizontal distances along transects are in kilometres. Source: I. Polyakov, International Arctic Research Center.

through the Bering Strait in 2007 was comparable (about 1 Sv) to the previous highest annual value; together with positive anomalies of heat content, it appears that the Bering Strait heat flux in 2007 was also at a record-length high (Dickson and Farbach, 2010: p. 5). In recent years, a more immediate driver of sea-ice melt appears to be associated with a near-surface temperature maximum (NSTM) at a depth of ~25 m in Canada Basin, where the NSTM may serve to maintain thinner ice during winter, an earlier melt during spring, and hence year-to-year persistence of ice anomalies (Jackson et al., 2010). The somewhat deeper (~60 m) layer of Pacific Summer Water (Figure 2.13) may also have played a role in the summer ice retreat. Thinner ice in this region could also favor the wind-stress and ice-loss feedback mechanism proposed by Shimada et al. (2006). Second, North Atlantic inflows appear to be increasingly consequential for the Arctic Ocean. North Atlantic inflows to the Arctic through the Barents Sea appear to have reached their highest temperatures in 100 years (Holliday et al., 2007). In addition, mooring measurements from the Arctic Ocean indicate the propagation of increasingly warm water in a cyclonic direction around the Arctic Ocean shelf break (Figure 2.14), consistent with the mean pattern of Arctic Ocean currents (Figure 2.13). While the Atlantic layer containing this anomalous heat is subducted several hundred metres below the surface as it circulates through the Arctic Ocean, the role of its anomalous heat in the sea-ice retreat and other cryospheric variations has not been firmly established. The Atlantic water heat influx to the Arctic Ocean appears to be characterized by increasingly warm pulses separated by brief respites (Polyakov et al., 2010). As shown in Figure 2.14, one such respite of cooling appears to have occurred in 2008/2009, although the Fram Strait cross-sections show that the next pulse of warming may now be poised to enter the Arctic Ocean. Finally, IPY measurements in various straits of northern Canada have confirmed the importance of the Canadian Archipelago for the freshwater flux from the Arctic Ocean to the North Atlantic, providing significant inputs to the Atlantic Meridional Overturning Circulation and thus to climate (Dickson and Fahrbach, 2010). The nature of the linkage between Arctic Ocean outflows and climate remains a holy grail of Arctic research.

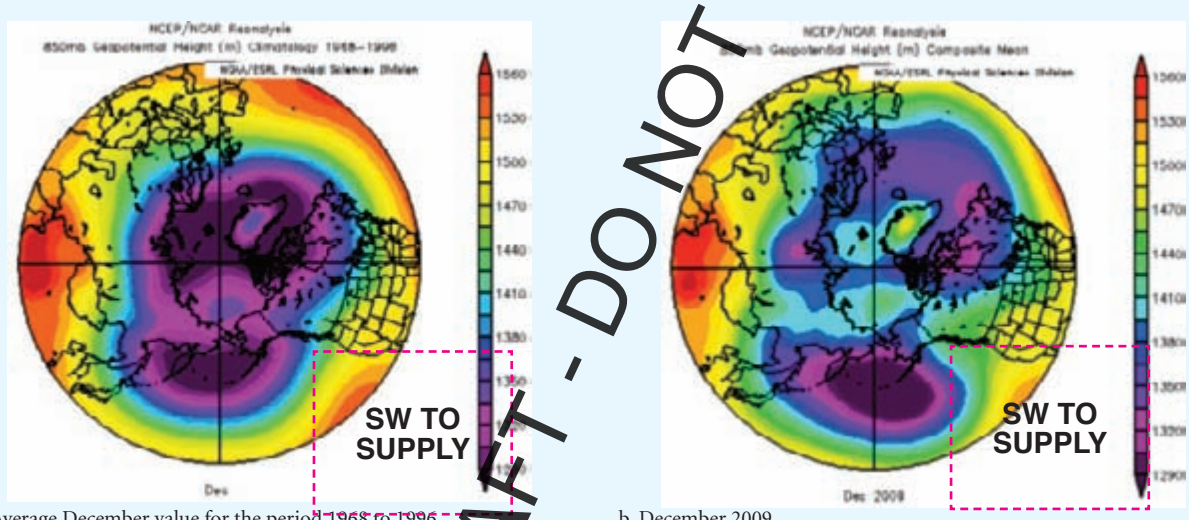
2.7. Conclusions

This review of recent Arctic climate variations, with an emphasis on air temperature and precipitation, serves two purposes. First, it shows that variations in Arctic surface air temperature and precipitation are key drivers of recent Arctic cryospheric change. Taken together with the cryospheric changes presented in later chapters of this report, a picture emerges of changes that are generally consistent across the Arctic system. Second, it shows that the Arctic is exhibiting markedly different behavior now than earlier in the instrumental period and, in the case of summer surface air temperatures, relative to 2000-year reconstructions of past variations. The available data do not permit determinations of whether other variables or seasons have experienced recent excursions outside their 2000-year range of variability. Nevertheless, the unprecedented warmth of the past five years reinforces the urgency of an assessment of cryospheric changes in the North. From a general

Box 2.1. Title needed

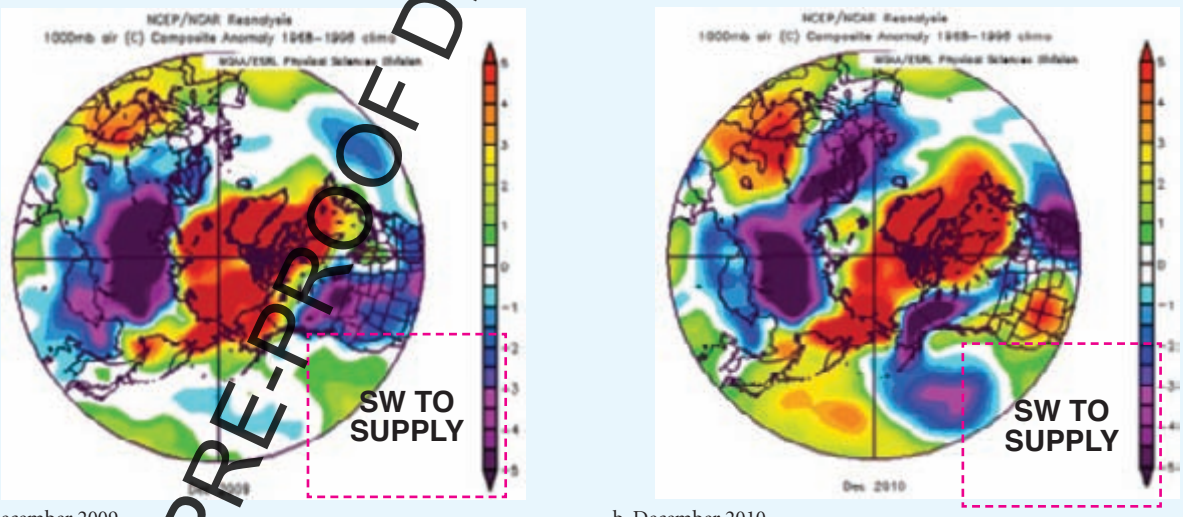
A recent Arctic influence on mid-latitude weather was the emergence of strong meridional atmospheric circulation in winter 2009/10 and the beginning of winter 2010/11, which allowed cold air to advect southward into eastern North America and Asia, and northern Europe (Cattiaux et al., 2010; L'Heureux et al., 2010; Seager et al., 2010). Cold air is normally trapped within the Arctic in winter by strong Polar Vortex winds, which circle the North Pole consistent with the low geopotential height field over the central Arctic (shown in purple in Figure 2.15(a). This pattern broke down in December 2009 (Figure 2.15b); the Polar Vortex winds, normally blowing from west to east, weakened as shown by the increased geopotential height field (greens) over the central Arctic, and north-south winds increased allowing cold Arctic air to spill southward. This created the Warm Arctic-Cold Continent Climate Pattern, shown in Figure 2.16 for December 2009. December 2010 had a similar pattern. Higher than normal Arctic temperatures (red) were seen especially in regions that were sea-ice-free in summer: north of Alaska and in the Barents Sea. The cold continents (purple) are seen where Arctic air penetrated southward. One indicator of a weak Polar Vortex is the North Atlantic Oscillation (NAO) index. Winter 2009/10 had the lowest NAO value in 145 years of historical record (www.cgd.ucar.edu/cas/jhurrell/indices.html).

Attribution for these cold mid-latitude winters is nearly impossible given the chaotic nature of atmospheric circulation. But given the extreme atmospheric circulation in back-to-back years and the recent changes in the Arctic, a possible weak Arctic–sub-Arctic linkage cannot be ruled out. Warmer Arctic air in autumn is less dense and increases the geopotential thickness between constant pressure surfaces, thus working against the stability of the Polar Vortex (Schweiger et al., 2008; Serreze et al., 2008; Overland and Wang, 2010). Models results also suggest Arctic–sub-Arctic teleconnections (Singarayer et al., 2006; Sokolova et al., 2007; Seierstad and Bader, 2008; Budikova, 2009; Deser et al., 2010; Kumar et al., 2010; Petoukhov and Semenov, 2010). Increased sub-Arctic weather variability seems a possibility as increasing amounts of sea ice are lost as the mid-century approaches.



a. Average December value for the period 1968 to 1996 b. December 2009

Figure 2.15. Arctic atmospheric pressure fields: normal 850 mb geopotential height values were observed in December from 1968 to 1996 (left) with unusual 850 mb geopotential height maximums observed over the Arctic in December 2009 (right). Figures from NOAA/ESRL Physical Sciences Division.



a. December 2009 b. December 2010

Figure 2.16. The Warm Arctic-Cold Continents Climate Pattern for (a) December 2009 and (b) December 2010. Plots show anomalies or deviations from the normal 1000 mb air temperature values which were observed from 1968 to 1996. Data are from the NCEP-NCAR Reanalysis through the NOAA/ESRL Physical Sciences Division.

climate perspective, the results point to the emergence of the ice-albedo-temperature feedback in the seasonal and spatial patterns of the recent surface air temperature anomalies in the Arctic. This is perhaps the most fundamentally important development in high-latitude climate since the Arctic Climate Impact Assessment in 2004 (ACIA, 2005). Given the absence of strong anomalies in large-scale circulation drivers such as the AO and the PDO in the past five years, the recent events support the changes that were anticipated in the ACIA report (ACIA, 2005) and echo the statement of Serreze and Francis (2006: p. 241) that "*Given the general consistency with model projections, we are likely near the threshold when absorption of solar radiation during summer limits ice growth the following autumn and winter, initiating a feedback leading to a substantial increase in Arctic Ocean surface air temperatures*". The cryospheric and atmospheric changes of the past five years indicate that this threshold may well have been crossed.

The results presented in this chapter also point to various observational needs. These are discussed in more detail in Chapter 11 (Section 11.6, Key et al.). Specifically, the precipitation estimates summarized in Section 2.3 are, by necessity, for Arctic land areas only. There are no systematically compiled sources of precipitation over the Arctic Ocean and its marginal seas, although it should be noted that Peterson et al. (2006) deduced recent increases in precipitation over the sub-Arctic North Atlantic based on computed moisture flux convergences in atmospheric re-analyses. Moreover, as described in Chapter 11 (Section 11.6, Key et al.), the station-derived estimates of precipitation for land areas in the Arctic have uncertainties arising from measurement errors (e.g., gauge undercatch of snow, for which only approximate correction procedures exist) and from the preferential siting of precipitation gauges in low-elevation areas. Temperatures over the Arctic Ocean are also subject to uncertainties, as the estimates of surface air temperatures over ice-covered seas are generally based on extrapolation of temperature anomalies from nearby land areas. Satellite-derived estimates of Arctic surface (skin) temperatures are generally biased toward cloud-free conditions. As atmospheric re-analyses for high-latitude regions are improved (e.g., Bromwich et al., 2010), estimates of pan-Arctic temperature as well as precipitation will become more robust.

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Chapter 3: Climate Model Projections for the Arctic

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Key Findings

- Most of the information on future changes in the Arctic is based on atmosphere-ice-ocean general circulation models from 18 climate modeling centers that formed part of the Intergovernmental Panel on Climate Change Fourth Assessment in 2007.

- No single model can be considered 'best' because results differ depending on the variable of interest (temperature, sea ice, sea-level pressure), location, and verification method, in apparently unsystematic ways.

- While details depend on location, time of year, model, and assumed future increases in greenhouse gas emissions, the consensus suggests a general increase in Arctic-wide autumn and winter surface air temperatures of 3 to 6 °C by 2080, a nearly sea-ice free September by 2050, and a general increase in precipitation over future decades.

- Owing to the random influence of natural variability, the single future trajectory that we will live through will differ somewhat from the smooth average of all possible future trajectories from the general circulation models. For example, the observed sea-ice loss in summer 2007 was greater than most trajectories projected by the general circulation models.

Summary

Projections of future changes in the cryosphere are dependent on interpretation of results from multiple climate model forecasts at regional scales. In setting the stage for discussing the various physical and biological components of the SWIPA project, this chapter surveys the major causes of the range of results provided by multiple atmosphere-ice-ocean general circulation models and recommends a methodology for summarizing the available projections in practical applications such as this assessment. Strategies are assessed for reducing the uncertainties from multiple model projections while acknowledging that multiple model simulation results (termed 'ensemble members') must be retained because forecasts should include a range of results caused by natural variability present in the real world and represented in the climate models. Results from international modeling centers form an 'ensemble of opportunity' from which composite projections can be made. However, experience suggests that these models need to be subjected to a selection methodology and independent interpretation.

Interpretation of results from 18 international modeling centers developed for the Fourth Assessment of the Intergovernmental Panel on Climate Change, suggests an increase in Arctic-wide autumn and winter surface air temperatures of 3 to 6 °C by 2080, a nearly sea-ice free September by 2050, and a general increase in precipitation. Loss of sea ice and increased temperatures are a coupled process with the greatest temperature increases seen in the autumn. The Arctic shows the largest future temperature changes on the planet.

Considerations that guide in the initial screening of a set of models are model resolution, comprehensiveness of processes included in the model, and sophistication of the parameterizations. A

main procedure for selecting a subset of the models is based on comparison to observations during the model hindcast period. The quality of individual model performance varies for different regions, variables, and evaluation metrics. For this reason, the use of a single model forecast is not recommended. The coarse resolution of most current climate models dictates caution in their application on smaller scales in heterogeneous regions such as along coastlines or in locations with rugged orography.

3.1. Introduction

Much of the information on future change in the Arctic contained in this assessment report is based on comprehensive atmosphere-ice-ocean general circulation models (GCMs) from 18 international climate modeling centers that were developed for the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) in 2007. As results from these models can vary, it is important to assess the robustness of the information provided by these models. The strategy for using these projections of change is an emerging research focus that has a direct bearing on the credibility of the future trajectories of cryospheric components. Accordingly, this chapter provides a review of climate model treatments of the Arctic, an assessment of their performance in the Arctic, and an overview of the key considerations that guide the use of these models in assessments of the likely impacts of future climate change.

The IPCC used projections from about two dozen GCMs developed by 18 international modeling centers in AR4 (Solomon et al., 2007). Table 3.1 lists the models and their resolution. A major feature of AR4 was that the detailed numerical information from these projections was made available to the wider scientific community for independent review; the projections and corresponding 20th century hindcasts are now archived as part of the Climate Model Intercomparison Project (CMIP3) by the Program for Climate Model Diagnosis and Intercomparison (PCMDI; www-pcmdi.llnl.gov). The model results have been evaluated independently by the authors of this chapter in addition to many other groups. While much of the earlier work emphasized global- and hemispheric-scale changes, governments, management agencies, and other stakeholders need regional predictions of possible future climate states in order to assess and plan for possible ecological and societal impacts and necessary adaptation actions. This chapter focuses on two central questions: how dependable are these model projections at regional scales and what is the limit of their utility?

Table 3.1. Selected model features. Salient features of the participating AR4 coupled models are listed by IPCC ID along with the calendar year of origin ('vintage'). Also listed are the respective sponsoring institutions, the horizontal and vertical resolution of the model atmosphere and ocean, the pressure of the atmospheric top, as well as the oceanic vertical coordinate (depth or density), and upper boundary condition (free surface or rigid lid). The three right-most columns list the characteristics of sea ice dynamics and structure (e.g., rheology vs. 'free drift' assumption and inclusion of ice leads), and whether adjustments of surface momentum, heat, or freshwater fluxes are applied in coupling the atmosphere, ocean, and sea ice components. Land features such as the representation of soil moisture (single-layer 'bucket' vs. multi-layered scheme) and the presence of a vegetation canopy or a river routing scheme also are noted. Relevant references describing details of the sea-ice and land surface formulations also are cited. Source: Randall et al. (2007).

Model ID, Vintage	Sponsor(s), Country	Atmosphere	Ocean	Sea ice	Coupling	Land
		Top Resolution	Resolution Z Coord., Top BC	Dynamics, Leads (Sources)	Flux Adjustments	Soil, Plants, Routing(Sources)
1: BCC-CM1, 2005	Beijing Climate Center, China	top = 25 hPa T63 (1.9°×1.9°) L16	1.9° × 1.9° L30 depth, free surface	no rheology or leads	heat, momentum	layers, canopy, routing
2: BCCR-BCM2.0, 2005	Bjerknes Centre for Climate Research, Norway	top = 10 hPa T63 (1.9° × 1.9°) L31	0.5 – 1.5° × 1.5° L35 density, free surface	rheology, leads (Hibler, 1979; Harder, 1996)	no adjustments	layers, canopy, routing (Mahfouf et al., 1995; Douville et al., 1995; Oki and Sud, 1998)
3: CCSM3, 2005	National Center for Atmospheric Research, USA	top = 2.2 hPa T85 (1.4° × 1.4°) L26	0.3 – 1° × 1° L40 depth, free surface	rheology, leads (Briegleb et al., 2004)	no adjustments	layers, canopy, routing (Oleson et al., 2004)
4: CGCM3.1(T47), 2005	Canadian Centre for Climate Modeling and Analysis, Canada	top = 1 hPa T47 (~2.8° × 2.8°) L31	1.9° × 1.9° L29 depth, rigid lid	rheology, leads (Hibler, 1979; Flato and Hibler, 1992)	heat, freshwater	layers, canopy, routing (Verseghy et al., 1993)
5: CGCM3.1(T63), 2005	Canadian Centre for Climate Modeling and Analysis, Canada	top = 1 hPa T63 (~1.9° × 1.9°) L31	0.9° × 1.4° L29 depth, rigid lid	rheology, leads (Hibler, 1979; Flato and Hibler, 1992)	heat, freshwater	layers, canopy, routing (Verseghy et al., 1993)
6: CNRM-CM3, 2004	Météo-France, Centre National de Recherches Météorologiques, France	top = 0.05 hPa T63 (~1.9° × 1.9°) L45	0.5 – 2° × 2° L31 depth, rigid lid	rheology, leads (Hunke and Dukowicz, 1997; Mélia, 2002)	no adjustments	layers, canopy, routing (Mahfouf et al., 1995; Douville et al., 1995; Oki and Sud, 1998)
7: CSIRO-MK3.0, 2001	CSIRO Atmospheric Research, Australia	top = 4.5 hPa T63 (~1.9° × 1.9°) L18	0.8° × 1.9° L31 depth, rigid lid	rheology, leads (O'Farrell, 1998)	no adjustments	layers, canopy (Gordon et al., 2002)
8: ECHAM5/MPI-OM, 2005	Max Planck Institute for Meteorology, Germany	top = 10 hPa T63 (~1.9° × 1.9°) L31	1.5° × 1.5° L40 depth, free surface	rheology, leads (Hibler, 1979; Semtner, 1976)	no adjustments	bucket, canopy, routing (Hagemann, 2002; Hagemann and Gates, 2001)
9: ECHO-G, 1999	Meteorological Institute of the University of Bonn, Meteorological Research Institute of KMA, and Model & Data Group,	top = 10 hPa T30 (~3.9° × 3.9°) L19	0.5 – 2.8° × 2.8° L20 depth, free surface	rheology, leads (Wolff et al., 1997)	heat, freshwater	bucket, canopy, routing (Roeckner et al., 1996; Dümenil and Todini, 1992)

	Germany/Korea					
10: FGOALS-g1.0, 2004	LASG/Institute of Atmospheric Physics, China	top = 2.2 hPa T42 (~2.8° × 2.8°) L26	1.0° x 1.0° L16 eta, free surface	rheology, leads (Briegleb et al., 2004)	no adjustments	layers, canopy, routing
11: GDFL-CM2.0, 2005	U.S. Dept. of Commerce, NOAA, Geophysical Fluid Dynamics Laboratory, USA	top = 3 hPa 2.0° × 2.5° L24	0.3 – 1.0° × 1.0° depth, free surface	rheology, leads (Winton, 2000; Delworth et al., 2006)	no adjustments	bucket, canopy, routing (Milly and Shmakin, 2002; GFDL GAMDT, 2004)
12: GDFL-CM2.1, 2005	U.S. Dept. of Commerce, NOAA, Geophysical Fluid Dynamics Laboratory, USA	top = 3 hPa 2.0° × 2.5° L24	0.3 – 1.0° × 1.0° depth, free surface	rheology, leads (Winton, 2000; Delworth et al., 2006)	no adjustments	bucket, canopy, routing (Milly and Shmakin, 2002; GFDL GAMDT, 2004)
13: GISS-AOM, 2004	NASA/Goddard Institute for Space Studies, USA	top = 10 hPa 3° × 4° L12	3° × 4° L16 mass/area, free sfc.	rheology, leads (Flato and Hibler, 1992)	no adjustments	layers, canopy, routing (Abramopoulos et al., 1988; Miller et al., 1994)
14: GISS-EH, 2004	NASA/Goddard Institute for Space Studies, USA	top = 0.1 hPa 4° × 5° L20	2° × 2° L16 density, free surface	rheology, leads (Liu et al., 2003; Schmidt et al., 2004)	no adjustments	layers, canopy, routing (Friend and Kiang, 2005)
15: GISS-ER, 2004	NASA/Goddard Institute for Space Studies, USA	top = 0.1 hPa 4° × 5° L20	4° x 5° L13 mass/area, free sfc.	rheology, leads (Liu et al., 2003; Schmidt et al., 2004)	no adjustments	layers, canopy, routing (Friend and Kiang, 2005)
16: INM-CM3.0, 2004	Institute for Numerical Mathematics, Russia	top = 10 hPa 4º × 5º L21	2º × 2.5º L33 sigma, rigid lid	no rheology or leads	regional freshwater	layers, canopy, no routing (Volodin and Lykosoff, 1998)
17: IPSL-CM4, 2005	Institut Pierre Simon Laplace, France	top = 4 hPa 2.5º × 3.75º L19	12º × 2º L31 depth, free surface	rheology, leads (Goosse and Fichefet, 1999)	no adjustments	layers, canopy, routing (Krinner et al., 2005)
18: MIROC3.2(hires), 2004	Center for Climate System Research (University of Tokyo), National Institute for Environmental Studies, and Frontier Research Center for Global Change (JAMSTEC), Japan	top = 40 km T106 (~1.1° × 1.1°) L56	0.2º × 0.3º L47 sigma/depth,free surface	rheology, leads (K-1 Developers, 2004)	no adjustments	layers, canopy, routing (K-1 Developers, 2004; Oki and Sud, 1998)
19: MIROC3.2(medres), 2004		top = 30 km T42 (~2.8º × 2.8º) L20	0.5 – 1.4º × 1.4º L43 sigma/depth, free surface	rheology, leads (K-1 Developers, 2004)	no adjustments	layers, canopy, routing (K-1 Developers, 2004; Oki and Sud, 1998)

20: MRI-CGCM2.3.2, 2003	Meteorological Research Institute, Japan	top = 0.4 hPa T42 (~2.8º × 2.8º) L30	0.5 – 2.0º × 2.5º L23 depth, rigid lid	free drift, leads (Mellor and Kantha, 1989)	heat, freshwater, momentum (12S–12N)	layers, canopy, routing (Sellers et al., 1986; Sato et al., 1989)
21: PCM, 1998	National Center for Atmospheric Research, USA	top = 2.2 hPa T42 (~2.8º × 2.8º) L26	0.5 – 0.7º × 1.1º L40 depth, free surface	rheology, leads (Hunke and Dukowicz 1997, 2003; Zhang et al., 1999)	no adjustments	layers, canopy, no routing (Bonan, 1998)
22: UKMO-HadCM3, 1997	Hadley Centre for Climate Prediction and Research/Met Office, UK	top = 5 hPa 2.5º × 3.8º L19	1.5º × 1.5º L20 depth, rigid lid	free drift, leads (Cattle and Crossley, 1995)	no adjustments	layers, canopy, routing (Cox et al., 1999)
23: UKMO-HadGEM, 2004	Hadley Centre for Climate Prediction and Research/Met Office, UK	top = 39.2 km ~1.3º × 1.9º L38	0.3 – 1.0º × 1.0º L40 depth, free surface	rheology, leads (Hunke and Dukowicz, 1997; Semtner, 1976)	no adjustments	layers, canopy, routing (Essery et al., 2001; Oki and Sud, 1998)

There are several arguments for suggesting that these models can provide reliable projections. Models are built on well-known physical principles, and many large-scale aspects of present-day climate are simulated quite well by these models (Randall et al., 2007; Knutti, 2008). Furthermore, biases in simulated climate by different models tend to be unsystematic (Raisanen, 2007), although there is some covariability of errors (Jun et al., 2008) and this statement is not valid for sea ice (Wang and Overland, 2009). While it might be thought that there would be considerable convergence between different models in their simulation results for the 20th century and projections for the 21st century, given that they are trying to simulate responses to similar forcing, there is in fact considerable variability in the ability of models to hindcast climate patterns based on location, variable of interest, and evaluation metrics (e.g., means, variance, trends), with some models performing well according to some criteria but not to others. Thus, the question of model reliability has no simple quantitative answer; there is no one best model (Gleckler et al., 2008). Contributing to these differences are model structure, physical parameterizations, and all the reasonable choices made during model development. The impact of natural variability, producing a range of results for similar model runs, is another challenge in comparing models with each other and with observational data. While practitioners are aware of many of these deficiencies, there is nonetheless a need to understand and make the best possible use of the output of existing models while improved models are being developed.

3.2. Projections of Arctic climate and key drivers of Arctic cryospheric change

Global climate models project that the Arctic will warm at a greater rate over the coming decades and century compared with other regions of the globe (Serreze and Francis, 2006; Overland, 2009). Subsequent chapters of this assessment report make use of 21st century projections of specific cryospheric variables (snow, sea ice, permafrost). While many factors contribute to changes over this time frame, atmospheric temperature and precipitation are among the key drivers that affect all cryospheric variables. Therefore, for these fields this chapter updates the Arctic Climate Impact Assessment (ACIA) projections of change on a seasonal basis. The update is based on the CMIP3 model simulations discussed in the existing literature.

While all models have biases, many of the biases tend to offset when the models' simulations are aggregated or composited. An important question is: how well do composites of the model simulations capture recent trends? This question can be answered by compositing the CMIP3 model simulations of the past half century and comparing the simulated trends with the observed trends, omitting the FGOALS and GISS-ER models (see Table 3.1) on the basis described in the following sections. Figure 3.1 shows the simulated change of annual mean Arctic temperature for 1957 to 2006, together with the temperature change from the National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR) re-analysis for the same period. The scaling (color bars) is identical for both the simulated and observed change maps. It is apparent that the CMIP3 composite captures the change in terms of (i) polar amplification, with maximum warming over the Arctic, and (ii) the magnitude of the warming, which averages between 0.5 and 1.0 °C over the domain in both cases, with local maxima of about 2 °C in each case. Not surprisingly, there are some differences in the regional details. For example, the re-analysis shows an area of cooling in the North Pacific, while the CMIP3 composite does not. Also, the location of the maximum warming is over the Beaufort Sea in the re-analysis and over the Barents Sea in the CMIP3 composite, indicating that some of the models demonstrate a loss of sea ice in the Barents Sea during 1957 to 2006. However, the CMIP3 model runs for this period were not set up to simulate observed regional variations, which may be generated from internal variability particularly originating from the oceans. It is also important to note that the agreement between the simulated and observed trends is weaker seasonally, especially during winter when the actual climate system has circulation-driven variations of temperature. Similarly, the individual models show much wider ranges of changes hemispherically and regionally because natural variability dominates a single realization over a 50-year period – as noted in Section 3.3 and as shown by Wang et al. (2007). Figure 3.2 shows that the 50-year temperature change for winter from three individual models displays large spatial and across-model

variability. The spatial scale of this variability is consistent with circulation-driven anomalies of temperature. The winter CMIP3 composite temperature change map smoothes much of the across-model variance (see Figure 3.2), but it does not – nor should it – match the spatial variations of the observed winter change, which have areas of cooling in eastern Siberia and near Greenland, probably due to the real atmosphere's natural variations during 1957 to 2006.

Fourteen of the global climate models used in the IPCC AR4 (Solomon et al., 2007) project warming for the Arctic beyond 2030 for each of three increasingly larger greenhouse gas emissions scenarios (labeled B1, A1B, and A2 according to IPCC terminology – see Box 3.1), although interannual variability of the surface air temperatures in the Arctic is large, especially over land (Chapman and Walsh, 2007). Projected changes vary widely among models and emissions scenarios. By the end of the 21st century, the projected annual mean temperature changes range from increases of +1.0 to +5.5 °C for the B1 scenario, increases from +2.5 to +7.0 °C for the A1B scenario, and increases from +4.0 to +9.0 °C for the A2 scenario. These ranges are larger than the corresponding values for the entire globe. Differences between the three emissions scenarios are projected to be small in the first half of the 21st century, but increase toward the end of the 21st century. Warming rates are smaller over land and largest over the Arctic Ocean. The greatest annual warming over land is near the Barents Sea (~6.5 °C for the A1B scenario). All models project the largest warming in autumn and winter, although the rates vary considerably among the models and across emissions scenarios. Walsh et al. (2009) evaluated the performance of fifteen models used in IPCC AR4 (Solomon et al., 2007) and found that models with smaller errors tended to simulate greater warming and greater increases in precipitation amount over the Arctic (60° to 90° N). Such findings have application for the selection of subsets of models to reduce the associated uncertainties, although there is no guarantee that the models that best capture recent climate will produce the best projections of future change.

Box 3.1. IPCC Emissions Scenarios

Emissions scenarios imply that model simulations of future climate conditions are made specifying different rates of increase in the emission of greenhouse gases. In IPCC AR4, the A2 emissions scenario (for a very heterogeneous world with high population growth, modest economic development and slow technological change) implies a greater carbon dioxide (CO_2) increase than the B2 emissions scenario (for a world with intermediate population and economic growth, emphasizing local solutions to economic, social, and environmental sustainability) and the A1B emissions scenario (a world of very rapid economic growth, a global population that peaks in mid-century and rapid introduction of new and more efficient technologies, with a balance across fossil intensive and non-fossil energy resources) that is an in-between case.

It is apparent from the composite CMIP3 projections of seasonal temperature change for the late 21st century (Figure 3.3) that warming dominates in all seasons, and that the warming is amplified compared to previous climatology in all seasons except in summer when the melting 'ice bath' of the Arctic Ocean constrains the surface air temperature to remain close to 0 °C. The annual mean warming is polar-amplified, in much the same way as the observed and simulated patterns of annual mean warming for the 1957 to 2006 period (Figure 3.1). The warming is largest over the Arctic in autumn and winter (3 to 6 °C by 2080), consistent with a loss of summer sea ice, greater absorption of solar radiation during summer, and release of this additional heat to the atmosphere during the cold season. Especially during winter, the spatial pattern bears the signature of sea ice reduction, with local maxima of 6 to 7 °C in the areas that have lost sea ice. Over much of the northern land areas, the warming is between 2 and 3 °C. It should be noted that the projections in Figure 3.3 are based on the A1B scenario. The corresponding spatial patterns for the A2 and B1 scenarios are similar to the A1B patterns shown in Figure 3.3, but the magnitudes of the warming are 30% to 50% larger for the A2 scenario and 30% to 50% smaller for the B1 scenario.

It is difficult to establish cause and effect between loss of sea ice and increased air temperature. Wang and Overland (2009) demonstrated that the more reliable models show a greater reduction rate in summer sea-ice loss (see Chapter 9). Further, given the major observed sea-ice reductions in 2007 to 2010, the faster sea-ice loss ensemble members should be favored (Holland et al., 2008), giving a nearly sea-ice free September by the mid-21st century.

Kattsov et al. (2007b) projected changes in precipitation amounts for 2041 to 2060 and 2080 to 2099 relative to measured precipitation data for 1980 to 1999 for a subset (13) of 21 of the IPCC AR4 GCMs forced by emissions scenarios B1, A1B, and A2. All of the models and scenarios showed increased precipitation across the Arctic through the 21st century, with much larger percentage increases than shown for the global mean precipitation and with distinct regional patterns. Percentage increases are generally largest at higher latitudes and most pronounced over northeast Greenland followed by coastal Siberia and the Canadian Arctic Archipelago. Percentage increases projected by 2080 to 2099 vary across the Arctic from a range of 5% to 40% (B1 scenario) to a range of 5% to 70% (A2 scenario), with the largest increases projected for northeast Greenland. The Arctic precipitation changes have a pronounced seasonality, with the largest relative increases in winter and autumn and the smallest in summer. The across-model scatter of the precipitation increase for each emissions scenario is large, but smaller than the scatter among different models in the baseline period (Kattsov et al., 2007b).

Figure 3.4 shows the CMIP3 composite projections of change in precipitation for summer and winter in the period 2070 to 2090. Although precipitation fields are inherently noisier than surface air temperature fields, compositing the simulations eliminates much of the spatial noise. The high latitudes are dominated by increases in precipitation in both seasons, with the largest increases found in coastal mountain ranges during winter. The only areas for which precipitation is projected to decrease are in the North Atlantic south of Greenland in winter and at the domain's periphery in summer when mid-latitude continental drying is projected by many models. The increase in precipitation is driven largely by changes in moisture flux convergence. Increases in Arctic precipitation are generally greatest in the models with the greatest warming. Only a proportion of the projected increases in precipitation will lead to increases in snow accumulation, since the fraction of precipitation falling as rain will increase under the projected increases in surface air temperature due to a general shortening of the period with sub-zero temperatures.

It should be emphasized that the projected increases in precipitation (typically 1 to 2 cm per season in the Arctic) do not preclude drying of the Arctic terrestrial regions. As illustrated in IPCC AR4 (Solomon et al., 2007: figure 10.3), increases in precipitation over Arctic land areas are accompanied by increases in evapotranspiration and runoff under the various greenhouse gas emissions scenarios, resulting in decreases in soil moisture. The net hydrological effect of a warming Arctic is an ongoing area of research, as precipitation and evapotranspiration are highly parameterized in models and hence subject to considerable uncertainties. Nevertheless, as reported in Chapter 2 there are suggestions of only small increases in Arctic precipitation over the past 50 to 60 years, while terrestrial drying has been sufficient to impact lake levels and fire frequencies in at least some Arctic terrestrial regions (GCCIUS, 2009).

3.3. Climate projection uncertainties

There are three main sources of uncertainty in the use of GCMs for climate projections: large natural variations (both forced and unforced), the range in emissions scenarios, and structural or model uncertainties. First, it is known that if climate models are run several times with slightly different initial conditions, the trajectory of day-to-day and indeed year-to-year evolution will have a different timing of events, even though the underlying statistical-spectral character of the 'model climate' tends to be similar for each run. This variability is a true feature of the climate system, and users of climate projections must recognize its importance. This uncertainty can affect decadal or even longer-term means, so it is highly relevant to the use of model-derived climate projections. To reduce the influence of the range of natural variability, the projections may be averaged over decades or, preferably, ensemble averages may be formed from a set of at least several model runs. Such a procedure can help distinguish the anthropogenic

contribution to climate change from the natural variability. There may also be interest in the possible changes in extreme states, which can result from a trend due to external (e.g., anthropogenic) effects combined with an episodic event due to intrinsic climate variability. Here, it is important to recognize the limitations in predicting the timing of events. Certainly, a drawback of considering only a single model run, rather than an ensemble of model runs, is that the intrinsic natural variability component cannot be clearly separated from the influence of anthropogenic external forcing.

A second source of uncertainty arises from the range in plausible emissions scenarios. Emissions scenarios are developed based on assumptions for the future development of humankind (Nakicenovic et al., 2000); they are converted into greenhouse gases and aerosol concentrations, which are then used to drive the GCMs in the form of external forcing specified in the CMIP3 models and summarized in IPCC AR4 (Solomon et al., 2007). Most of the CMIP3 models have made projections under the emissions scenarios A2, A1B, and B1 (see Box 3.1). Because the residence time of carbon in the atmospheric system is of the order of centuries, climate projections are relatively insensitive to the precise details of which future emissions scenarios are used over the next few decades, as the impacts of the scenarios are rather similar before the mid-21st century. However, for the latter half of the 21st century, and especially by 2100, the choice of emissions scenario dominates over natural variability and model-to-model differences, becoming the major source of uncertainty of climate projections (Solomon et al., 2007). With 2030 to 2050 as a timescale of interest, a single mid-range anthropogenic emissions scenario is often used (the A1B or the A1B and A2 together to increase the number of potential ensemble members), as their CO_2 trajectories are similar before 2050. However, some differences do emerge in the Arctic even before the mid-21st century as a result of, for example, the different scenarios of aerosols with their pronounced spatial distributions.

The third source of uncertainty is termed 'structural uncertainty', and is used here to refer not only to a model's configuration (resolution, vertical coordinate, grid point vs. spectral) but also to its formulation features. Different numerical approximations of the model equations, including spatial resolution, introduce part of the structural uncertainty. Sub-grid scale processes must be parameterized in all models; by necessity, these parameterizations are simplifications of complex processes and require tuning. The selection of parameterization schemes, and the tuning of their coefficients, is at least somewhat subjective, and the various modeling centers have made different choices. Multi-model ensembles have the advantage of sampling this structural uncertainty. For example, there is a minimum error when the composite contains five to seven models (Figure 3.5) when sampling the dependence of the root mean square error of Arctic surface air temperature on the number of highest-ranking models included in a multi-model ensemble composite. Apparently, single models are subject to structural errors, while as the number of models included in the average exceeds five to seven, the inclusion of additional, poorer-performing models compared to observations degrades the multi-model ensemble composite. This finding appears to be robust across variables and regions within the extratropical Northern Hemisphere, although its dependence on the length of the validation period requires further investigation.

3.4. Treatment of sea ice, snow, permafrost, glaciers, and ice sheets in climate models

In the treatment of the cryosphere, one of the major developments over the past several years has been the implementation of sea-ice dynamics in all coupled models. Previously, many models used very simple descriptions of sea ice or even none at all (Houghton et al., 2001). Table 3.1 contains short descriptions of the sea-ice treatment in each of the 23 participating models. Moreover, Table 3.1 shows that very few models employ artificial flux adjustment (regressing the solution back toward a base state) between the ocean (including sea ice) and the atmospheric component. Although a questionable modeling methodology improvement, the implications for model performance are less promising insofar as systematic biases are pronounced in many models, as the purpose of flux adjustments was to tie the models closer to observations. This is particularly evident in the Arctic, due to the complicated interplay between ocean, sea ice, land, and atmosphere.

Sea-ice components of current GCMs usually predict ice thickness (or volume), fractional cover, snow depth, surface and internal temperatures (or energy), and horizontal velocity. Some models now include prognostic sea-ice salinity (Schmidt et al., 2004). Sea-ice albedo is typically prescribed, with only crude dependence on ice thickness, snow cover, and puddling effects. The complexity of sea-ice dynamics varies from the relatively simple 'cavitating fluid' model (Flato and Hibler, 1992) to the viscous-plastic model (Hibler, 1979), which is computationally expensive, particularly for global climate simulations. The elastic-viscous-plastic model (Hunke and Dukowicz, 1997) is being increasingly employed, particularly due to its efficiency for parallel computers. Treatment of sea-ice thermodynamics typically includes constant conductivity and heat capacities for ice and snow (if represented), a heat reservoir simulating the effect of brine pockets in the ice, and several layers, the upper one representing snow. Some models include snow-ice formation, which occurs when an ice floe is submerged by the weight of the overlying snow cover and the flooded snow layer refreezes. In a significant advance over many models of a decade ago, many sea-ice models (even with fine resolution) incorporate sub-grid scale ice thickness distributions with several thickness 'categories', rather than considering the ice as a uniform slab with inclusions of open water. Although parameterizations of ridging mechanics and their relationship with the ice thickness distribution have improved (e.g., Toyota et al., 2004), inclusion of advanced ridging parameterizations has lagged other aspects of sea ice dynamics (rheology, in particular).

Land surface processes are also treated interactively in the models, although the processes represented vary widely among the models. A common feature in all of the more recent models is the implementation of river routing schemes, which allows for the return of surface waters back to the ocean. Cold land processes have received considerable attention with multi-layer snowpack models (e.g., Oleson et al., 2004) in many GCMs, as has the inclusion of soil freezing and thawing (e.g., Boone et al., 2000). Sub-grid scale snow parameterizations (Liston, 2004), snow-vegetation interactions, and the wind redistribution of snow (Essery and Pomeroy, 2004) are also more common in current models. A key feature of snow formulations is the parameterization of snow albedo, which depends on snow depth, age, and vegetative masking. Various model experiments in recent years have shown that the degree of vegetative masking of snow is important for climate locally and, in some experiments, remotely. The sophistication of the snow albedo parameterization, both on land and over sea ice, varies widely among models. A recent advance is the coupling of groundwater models into land surface schemes (e.g., Liang et al., 2003). These have only been evaluated locally but may be adaptable to global scales.

Below ground, a few layers typically are used to describe the diffusion of heat and moisture processes. In all cases, these layers are reaching no deeper than a few metres below the surface. Some models use a simple bucket description of soil moisture processes, while others use a more consistent interaction between moisture and energy fluxes, including freezing and thawing within the soils. Some details regarding these schemes are summarized as key words and listed in Table 3.1. A common feature of most GCMs is the treatment of the lower boundary condition as zero flux. This limits the ability of any long-term penetration of a climate change warming signal into the ground and instead causes a spurious reflection at the lower boundary, resulting in an exacerbated warming of the model soil layers. In cold climate regions, this poses a challenge as even the penetration of the annual signal of the temperature to depth is corrupted this way. Recent model developments are introducing more soil layers and a greater depth, resulting in substantial improvements in simulations of the subsurface annual cycle. Therefore, the permafrost simulations described in Chapter 5 are based on 'offline' simulations, i.e., a spatially distributed permafrost model with parameterizations, resolution, and depth that are enhanced relative to the terrestrial modules in GCMs.

A related model shortcoming has been a general ignorance of wetlands and organic soils, which are present in many Arctic regions, dominating the landscape in some areas. Shallow lakes, for example, cover as much as 30% of tundra areas such as the coastal plain of northern Alaska. High-latitude organic soils are included in some models (Wang et al., 2002). Regional climate model simulations also appear to improve when these special landscape types are taken into consideration.

When it comes to the treatment of glaciers and ice sheets, current GCMs are generally lacking in that the models do not include many processes important to glaciers, ice sheets, and their coupling to the atmosphere (and ocean). Glaciers are mostly of a size that is not resolved by GCMs and therefore are not treated interactively. Ice sheet models are used in calculations of long-term warming and sea-level scenarios, although they have not been incorporated into any of the 23 AR4 GCMs. The ice sheet models are run offline, and that is the approach followed in Chapter 8 for Greenland as well as for the smaller ice caps and glaciers. Ice sheet models are included in some Earth System Models of Intermediate Complexity (EMICs). However, the timescale of projected melting of the Greenland Ice Sheet may be different in coupled and offline simulations. One aspect of concern would seem to be the apparent discrepancy between the simplistic treatments of an ice sheet in the climate model, from which the output fields are then subsequently used to force an ice sheet model. The consequences of such an inconsistency still have to be assessed in the context of an evolving climate. Similarly, the approach used for assessing changes in permafrost conditions is based on offline modeling of permafrost conditions. With the limitations of shallow soil layers and lack of organic soil treatment in GCMs, this inconsistency in the driving climate needs to be kept in mind when using an offline approach such as has been employed in the present assessment.

3.5. Key considerations in model selection

There is no universal methodology for using multiple GCMs for climate projections. The performance of CMIP3 models varies with location, variable of interest, and analysis method, with some models performing well for some criteria but not for others (Gleckler et al., 2008; Walsh et al., 2009). If the climate impact problem is defined by needing a projection of a specific variable in a specific location, it is unclear whether it is preferable to base model selection decisions on the accuracy of the models for this local variable in hindcast simulations or with respect to overall measures of performance over a range of variables and larger regions. For example, advective processes can strongly affect local climate, implying that credible simulation of climate over a wider area will be important for changes at a particular location. Moreover, the coarse resolution of most current climate models certainly dictates caution in their application on smaller scales in heterogeneous regions such as along coastlines or in locations with rugged orography. Experience of conducting model evaluations over the past few years indicates the value of multiple and complementary approaches, i.e., models should be compared using different metrics. Due to the need to account for structural uncertainty, a single 'best' model should not be relied upon. Rather, a reduced group of models (a minimum of about five) may generally be optimal, but this may also be variable and regionally dependent. This approach contrasts with that of IPCC AR4, where averages were often presented across all contributed models of opportunity, regardless of quality.

There are three general concepts pertaining to multiple model evaluation. The first consideration is the quality of the underlying physics. Does the model include reasonable parameterizations of the physics of interest? For example, sea ice is represented in separate models at considerably different levels of sophistication. The number of layers in the ice and the presence or absence of ice transport, deformation, and thickness categories vary among models, as noted in Section 3.4. Similarly, if projections of permafrost are to be obtained directly from a GCM, the model should include processes affecting soil temperature and water in a manner that allows for credible simulations of temporal evolution. While this initial 'pre-screening' contains some inevitable subjectivity, it will leave the user with a set of model simulations that represents a more justifiable 'ensemble of opportunity'.

The second consideration is consistency: do different models in the ensemble of opportunity produce similar simulations? The degree of similarity can reflect how well underlying mechanisms are represented in the models. The third consideration is the skill of the models' hindcasts relative to various datasets or fields (e.g., for the 20th century). While past skill would seem to be a necessary condition for using a model, this provides no guarantee of accurate model projections under new external forcing or climate states (Reifen and Toumi, 2009). It is suggested that past performance should represent a primary criterion

in rating the models, but that hindcasts should not be the only consideration in choosing models on which to base a projection.

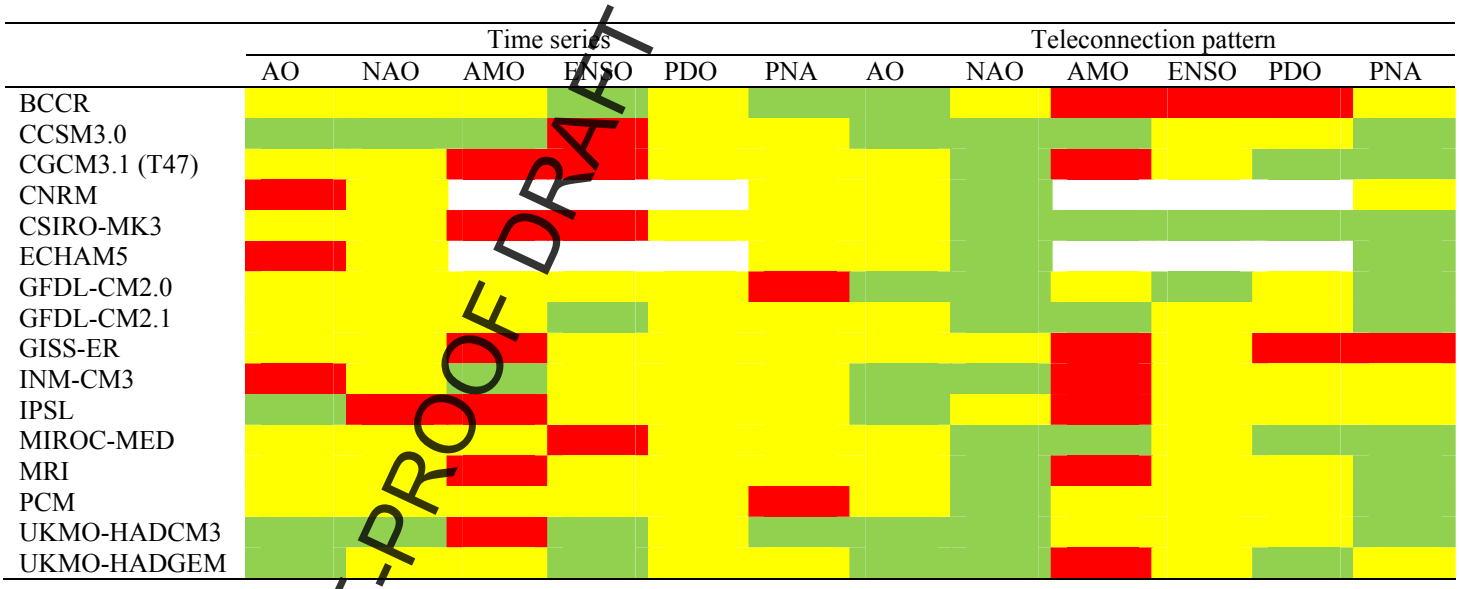
3.5.1. Choices for evaluating hindcast skill

For evaluating hindcast skill, choices include the selection of observational datasets, variables, regions, metrics, and weighting approaches. The metrics can include comparisons to mean climate, climate variability as represented by the annual cycle or interannual variability, correlation, or trends due to externally forced change.

Validation of models generally includes comparison to means. However, this approach has two drawbacks in relation to projections. First, the quality of projections should also relate to potential changes (or sensitivities) in the force balance or heat budget in the system, in addition to mean states. Second, models are tuned to represent the mean state and are not independent of the observational fields. However, if model projections depend on the state of the system (e.g., the state of sea ice) at the end of the 20th century, then having the correct mean conditions is critical (Wang and Overland, 2009).

Several authors recommended using the seasonal cycle for comparison, which appears to be a good choice as the range of radiative forcing and other conditions over the year is comparable to or greater than the magnitude of potential future climate shifts (Hall and Qu, 2006; Knutti et al., 2008; Walsh et al., 2009). It is also important to consider interannual variability ('variance'). For example, Stoner et al. (2009) documented the ability of models to capture the timescales and spatial patterns of the major modes of observed natural variability. Namely, the Arctic Oscillation (AO), the North Atlantic Oscillation (NAO), the Atlantic Multi-decadal Oscillation (AMO), the El Niño / Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO) and the Pacific North American (PNA), as shown in Table 3.2.

Table 3.2. Results of an evaluation by Stoner et al. (2009) of global climate model simulations of modes of natural variability. Green represents a good match, yellow a partial match, and red a lack of correspondence.



Using the correlation between model and observation time series is a poor choice for evaluating hindcast skill as the observations represent only a single realization of a system with strong natural variability. Comparisons must therefore utilize climate statistics rather than year-to-year values that manifest natural variability (i.e., chaotic uncertainty) (Kattsov et al., 2007a; Overland and Wang, 2007).

Using trends to evaluate hindcast skill would be another seemingly good choice, as the interest is in projecting future trends. However, over the past 20 to 50 years, internal variability is large and obscures the underlying trend due to external causes in the observation set, especially in the case of regional trends. In addition, a regional trend may not necessarily be attributable to an external driver. These problems compound the difficulty in using trends as a constraint.

Applying aggregate metrics (i.e., using multiple criteria) to model selection, while intellectually appealing, also has limitations. If an increasing number of variables or subregions are included as independent constraints, then an increasing number of models will exhibit deficiencies that make them candidates for exclusion. Experience suggests that a practical limit to model selection is rapidly reached for CMIP3 archived models. Likewise, if a 'model climate performance index' is developed by averaging performance rankings over many variables, it can produce inconsistencies. For example, what is the meaning of an additive index where a given model performs well on one variable but poorly on another? Is it better that a model does a reasonable job on a single or a few variables, or is middle-of-the-road performance over multiple variables more appropriate? User priorities may dictate the weighting of a particular variable in model evaluation, but deficiencies in fields of other variables may point to structural shortcomings that adversely affect a model's projections of change.

It is thus necessary to make choices regarding best practices selection methods. Inevitably, this leads to subjectivity in the approach for assessing the overall best models. In fact, the process should be viewed as reducing the impact of models with large hindcast errors rather than the selection of best models, while retaining several models as a measure of structural uncertainty. The 'best' multi-model ensemble will vary from application to application, depending on the particular priorities of the users. By necessity, credibility is increased by consensus between different approaches to selection and by the openness, simplicity, and transparency of the process. Careful documentation of the model selection procedure is essential. Although averaging is often a better procedure than simply using one model (Reichler and Kim, 2008), including all CMIP3 models as a grand average or a distribution is not a viable option, as some are obvious outliers with respect to observational correspondence. Thus, the following is noted:

1. An initial screening based on the underlying physics of various models, particularly as it pertains to the application at hand, will ensure that any quantitative evaluations are based on a set of models that is at least potentially credible.

2. In many applications, it will be advisable to eliminate the models that seriously fail to meet one or more observational constraints, based on actual or synthetic data (re-analyses). For example, do the selected variables have reasonable means and seasonal cycles (variance) close to the observations at a continental scale or above? This does not guarantee future performance, but it at least indicates that the projections are built upon reasonable background and users have more confidence in the models. It bears noting that even reproducing mean quantities approximately correct is different, as the CMIP3 models are initialized in the 19th century.

3. After eliminating poorly performing outlier models, the remaining subset of models can be evaluated for individual variables and regions of interest. The present analyses of GCM hindcast simulations show that different models perform differently in different regions and for different variables within a region, often without obvious reasons. Nevertheless, for specific applications, it is plausible that the models with better hindcast simulations of externally driven changes will also provide projections that are more reliable.

4. As a general rule, it is advisable to use multi-model ensembles to account for structural uncertainty. This is a sampling problem so a sample size of at least several models is desirable (Figure 3.5). A multi-model mean should tend to be more reliable than any individual model, even if the individual model is of high quality based on its performance in a 20th century hindcast. All current generation models include some non-systematic errors, and these errors can be reduced through averaging

(Reichler and Kim, 2008). Models selected by multivariable metrics generally outperform any individual model. As noted earlier, natural variability over the evaluation period also argues for the use of multiple models in obtaining a projection.

3.6. Meta-analysis

A meta-analysis approach is illustrated here by synthesizing the results from several earlier papers (Table 3.3). Walsh et al. (2009) analyzed surface air temperature (SAT) and sea-level pressure (SLP) north of 60° N using the observational constraint of the root mean square (RMS) error and variance over the seasonal cycle. Similarly, Gleckler et al. (2008) compared modeled mean values of sea-level pressure and 850 hPa temperatures (T850) with re-analyses in a complex model climate performance index. Wang et al. (2007) used the criteria of interannual variance of winter Arctic land surface air temperatures. Summer Arctic sea ice was investigated by Wang and Overland (2009), who required that the models not only be able to simulate the mean minimum sea-ice extent but also the magnitude of seasonality (March minus September extent). All of these evaluations involved assessing how well the model hindcasts for the 20th century matched observations. For reference, two more studies are also included, those of Reichler and Kim (2008) and Wu et al. (2008); the first included criteria for the entire planet, and the second concerned model sensitivity.

Table 3.3. Summary of CMIP3 model evaluations over the Northern Hemisphere and Arctic for selected variables.

	Region		Global	NH 20° to 90° N			NH 20° to 90° N		NH 60° to 90° N			NH 40° to 90° N		Global
	Source		Reichler and Kim, 2008	Gleckler et al., 2008			Walsh et al., 2009		Walsh et al., 2009		Wang et al., 2007	Overland and Wang, 2007	Wang and Overland, 2009	Wu et al., 2008
	Variables		14 variables	SAT	SLP	T850	SAT	SLP	SAT	SLP	SAT	Sea ice		Climate sensitivity
	Model ID	Country	Variance (annual mean)	Annual mean RMS error			RMS error[1] (monthly mean)		RMS error[1] (monthly mean)		Variance (winter)	Summer mean	Seasonal cycle	
1	BCCR-BCM2.0	Norway	16		14	red						fail	fail	
2	CCM3	USA	7	9		blue	2/2	13/13	2/3	15/4	pass	pass	pass	2.7
3	CGCM3.1(T47)	Canada	5	6	5	blue	10/10	4/5	11/8	2/4	pass	pass	fail	3.4
4	CGCM3.1(T63)	Canada	8	8	4	blue					pass	pass	fail	3.4
5	CNRM-CM3	France	14	13	7	blue	5/6	11/3	5/6	12/3	pass	pass	pass	
6	CSIRO-Mk3.0	Australia	13	7	19	red	12/13	9/10	14/13	9/14	pass	fail	fail	3.1
7	ECHAM5/MPI	Germany	3	2	1	blue	1/1	1/1	1/1	1/1	pass	pass	fail	3.4
8	ECHO-G(MIUB)	Germany/Korea			3						pass	pass	pass	3.2
9	FGOALS-g1.0(IAP)	China	12	20	16	red	15/15	10/9	15/15	13/15	fail			
10	GFDL-CM2.0	USA	10	11	9	red	14/4	4/4	9/5	8/10	pass	pass	fail	2.9
11	GFDL-CM2.1	USA	1	4	2	red	5/3	2/2	3/2	4/2	pass	fail	fail	3.4
12	GISS-AOM	USA	19	17	13	blue					fail	pass	fail	2.7
13	GISS-EH	USA	22	15	18	red					fail			
14	GISS-ER	USA	18	18	20	red	10/12	14/14	10/10	14/11	fail	fail	fail	
15	Ingv	Italy	11									fail	fail	
16	INM-CM3.0	Russia	15	14	12	red	10/11	9/11	6/7	7/9	pass	fail	fail	2.1
17	IPSL-CM4	France	17	19	21	red	12/8	15/15	7/9	11/13	fail	pass	pass	4.4
18	MIROC3.2(hires)	Japan	9	5	11	blue					fail	fail	fail	4.3
19	MIROC3.2(lowres)	Japan	12	10	8	blue	3/5	5/6	4/4	3/8	fail	pass	pass	4.0
20	MRI-CGCM2.3.2	Japan	4	12	15	red	7/7	6/7	13/12	11/12	fail	fail	fail	3.2
21	PCM	USA	21	16	17	red	14/14	12/11	13/14	5/7	pass			2.1
22	UKMO-HadCM3	UK	6	3	10	blue	6/10	7/8	8/11	6/6	pass	fail	fail	3.3
23	UKMO-HadGem1	UK	2	1	6	blue						pass	pass	4.4
24	BCC-CM1	China												
25	CSIRO-MK3.5	Australia											fail	

[1]The two numbers in these columns indicate the order of ranking for each model of selected variables for given region by papers published in the past – the first number is based on the original output, the second is when the bias has been removed; [2] 'Blue' indicates cases where a model fares better than the typical model with respect to the reference data, and 'red' indicates the contrary; [3] 'pass' and 'fail' indicate where models passed or did not pass the selection criterion.

With the high-latitude Northern Hemisphere and the SWIPA processes being the primary focus of this assessment, the meta-analysis ranks the models based on their performance on selected variables (surface air temperature, sea-ice extent, and sea-level pressure) over the Arctic and Northern Hemisphere when compared with observations. Of the CMIP3 models, some consistently perform better than others across the criteria. These are (in alphabetic order): CCSM3, CNRM-CM3, ECHO-G/MIUB, UKMO-HadGEM1 (shaded in purple in Table 3.3), which pass selection criteria for both Arctic surface air temperature and sea ice (Wang et al., 2007; Wang and Overland, 2009) and ranked relatively high by other studies (see also Table 3.3), as well as those shaded in yellow: CGCM3.1 (T47), CGCM3.1 (T63), ECHAM5/MPI, GFDL-CM2.1, MIROC3.2(medres), and UKMO-HadCM3. The second group of models ranked high generally, but fail at least one of the sea-ice extent or Arctic surface air temperature tests. The third group (shaded in green) includes only one model (IPSL-CM4), which performs well in the sea ice simulation but ranked generally low in all other studies.

Although there is consistency among the difference methods of ranking the models, there is also a caveat in the process. For example, the CCSM3 model performs relatively well for temperature and sea ice simulations but is biased for sea-level pressure in the Northern Hemisphere, as pointed out by Walsh et al. (2009). This is especially problematic over the Arctic region because when the sea-level pressure bias is removed, the model ranking is advanced from fifteenth place to fourth place (see Table 3.3). The issue is whether the models' bias should be removed before they are ranked?

It should be noted that this chapter does not endorse any particular model result; Table 3.3 is for illustrative purposes only. Unfortunately, the review in this chapter found no systematic conclusions for the differences in model projections. As noted again in the next sections, evaluation results differ based on location and variable of interest.

3.7. Two regional examples

3.7.1. Alaska

According to Walsh et al. (2009), the top-performing models for Alaska are the GFDL-CM2.1, ECHAM5/MPI-OM, CNRM-CM3, UKMO-HadCM3, and the Japanese MIROC3.2(medres), all of which belong to the first or second group of meta-rankings in Table 3.3. It should be noted that the ranking depends on whether the annual mean biases are included in the root mean square errors. The ECHAM5/MPI model ranks well below the median (thirteenth out of the fifteen models) in its simulation of Alaskan temperature primarily because the temperatures projected by the model are too cold throughout the year over Alaska (Figure 3.6). However, the amplitude of its seasonal cycle is very close to the observed, so removal of the annual mean bias improves the model's rank to the first of the same set of fifteen models. Hence, an important consideration in the use of GCM output is the nature of any bias adjustment that may be imposed. This example illustrates the important point that a climate model that simulates several variables well can be a relatively poor performer in the simulation of other variables, particularly on a regional basis. The converse is also true.

With regard to the reasons for the different levels of skill over Alaska and Greenland as well as the larger domains, no systematic relationship to model resolution has emerged. The models with the smallest RMS errors (ECHAM5/MPI and GFDL-CM 2.1) have resolutions that are neither the highest nor the lowest of the fifteen models. There is also no obvious relationship between model performance and the type of sea ice formulation in the models. Other candidates for explanations of the differences in model performance include the cloud formulations and radiative properties, the planetary boundary layer parameterization, and the land surface schemes of the various models (e.g., Kattsov et al., 2007b; Sorteberg et al., 2007). Biases in the large-scale atmospheric circulation, perhaps driven by processes outside the Arctic, are also candidates to explain the across-model differences in temperature, precipitation, and sea-level pressure.

An interesting result that emerged from this regional evaluation is that the top-performing models tended to be more sensitive to greenhouse forcing than the poorer-performing models. This similarity is apparent in the late 21st-century changes in Arctic surface air temperature, precipitation, and sea level pressure. Perhaps models that best capture the seasonal cycle of radiative forcing are more sensitive to external (greenhouse) forcing. The reasons for this correspondence require further investigation.

3.7.2. Sea ice simulation for the Barents and Bering Seas

The Bering Sea and Chukchi Sea between Alaska and northeastern Russia, and the Barents Sea north of Norway are known as seasonal sea-ice zones, where the sea ice advances in winter and spring and retreats in summer, leaving regions of open water. The climatology of sea-ice extents for the Barents Sea (65° to 82° N, 15° to 60° E) represents an exception to the statement in the introduction that biases in simulated climate by different models tend to be unsystematic. The simulated seasonal cycle of sea-ice conditions averaged over the period 1980 to 1999 shows that most models have too much sea ice in this region compared to the observed climatology (Figure 3.7). Expansion of sea ice in the Barents Sea is generally countered by ocean advection of heat into the region from the North Atlantic. An ongoing assessment indicates that the current-generation models in CMIP3 still have difficulty in achieving satisfactory simulations of ocean currents in this region. This over-specification of sea ice at the end of the 20th century had unfortunate implications for the full IPCC AR4. Too much sea ice in too many models meant that the simulated winter temperatures for the region during 1980 to 1999 were also too low (Figure 3.6). When the temperature change was projected for 2090 to 2999 relative to 1980 to 1999 and averaged over all the models, the changes were some of the largest on the planet, greater than 5 °C (Solomon et al., 2007: figures SPM6). The reason for this was not that this region would be excessively warm in the decade of 2090, but that the reference temperatures were incorrectly too low. While the average sea-ice extents are too high for the Barents Sea, several models (e.g., CCSM3, ECHAM5/MPI-OM, and MRI) do considerably well for the seasonal cycle of sea-ice extent, suggesting that the Barents Sea is a major candidate for a reduced set of projections based on model selection. However, since only one (CCSM3) of the three models mentioned above shows good agreement with the observations on Arctic-wide sea-ice extent (Wang and Overland, 2009), there is not enough confidence in a multi-model approach to the projection of future sea-ice conditions over the Barents Sea.

For the Bering Sea, the multi-step strategy has been used to evaluate model performance and select models suitable for regional projections (see procedure proposed in Section 3.5). For example, for the eastern Bering Sea (54° to 66° N, 175° to 155° W), the six models identified by Wang and Overland (2009) were initially selected, as these have simulated Arctic-wide sea-ice extents both in terms of the summer mean and the magnitude of seasonal cycle reasonably well. The models were then required be able to simulate the spring (April and May) sea-ice extent over the eastern Bering Sea with a less than 20% error from the observed value. The process led to four best-performing models over the eastern Bering Sea: CCSM3, CNRM-CM3, ECHO-G, and MIROC(medres). When a similar strategy was applied to the Chukchi Sea and the Beaufort Sea, the result was that all six models (CCSM3, CNRM-CM3, ECHO-G, IPSL-CM4, MIROC(medres), and UKMO-HadGEM1) identified by Wang and Overland (2009) passed the selection criteria. However, for the western Bering Sea and the Bering Strait, only one model (CCSM3) passed the selection criteria. For the Sea of Okhotsk, none of the models passed the selection criteria (Wang and Overland, 2009). This raises the question as to why some models perform better in one region than another, even for the same variable. Unfortunately, there is no clear answer to this at present.

3.8. Downscaling techniques

Direct use of output from GCMs for many cryospheric projections is not currently feasible due to the biases in model data on regional scales and due to the inability of models to resolve features of sea ice (e.g., polynyas and leads), glaciers, and topographic features that have major influences on snow and permafrost. In fact, most cryospheric features vary over smaller scales than climate model grid boxes, and terrestrial snow and ice often occupy complex terrain that is only coarsely resolved in the models'

underlying topography. The accumulation and ablation of ice are particularly sensitive to biases in air temperature, which controls the energy available for melt and the snow-rain ratio of precipitation. Hence, some method of downscaling is required that transfers global-scale climate information to local scales prior to making cryospheric projections. Downscaling considerations are discussed in the following paragraphs, with some examples from glaciology. Because the central issues are similar across cryospheric variables, the discussion is also relevant to other components of the cryosphere.

Statistical downscaling is often applied by establishing statistical relationships between meteorological quantities determined at the GCM and local scales, either using field observations or other suitable meteorological data such as data from climate re-analyses. In an application to glaciers, Radić and Hock (2006) applied 'local scaling' to correct for the biases in climate model air temperature and precipitation for projecting the mass balance of the Storglaciären glacier up to 2100. Downscaled temperature series were produced from GCMs and regional climate models (RCMs) by shifting the series by the averaged monthly differences between climate-model and local-scale data (ERA-40 re-analysis data, in this case) over a baseline period for which both GCM and local-scale data were available. Hence, the average seasonal cycle from ERA-40 was used as a reference by which seasonal cycles from the climate model were 'corrected'. The results highlight the importance of including the seasonally varying biases instead of assuming a constant bias throughout the year. Precipitation was scaled by the ratio of precipitation in the reference dataset (ERA-40) summed over the baseline period and the corresponding climate model precipitation sum. A drawback of statistical downscaling is the inherent assumption that the statistical relationships established over a baseline period will continue to hold in future climates.

An alternative approach to statistical downscaling is to use changes in GCM variables between defined future time intervals and a baseline period. These changes are then used to perturb observed local climate data to drive a mass balance model and project future mass balance changes. The changes in climate variables are often simply linearly interpolated between time intervals to allow for transient simulations. Due to generally large interannual variability, projected climate variable changes can be sensitive to the choice of baseline period. This is especially true when large fluctuations occur around the reference period or when the climate variable shows a trend during this period (Adalgeirsdottir et al., 2006).

Zhang et al. (2007) used a second downscaling approach, dynamical downscaling, employing the high-resolution Arctic MM5 regional atmospheric model driven by a global atmospheric re-analysis to obtain temperature and precipitation data on a 10-km resolution grid to force a glacier mass balance model. The results of mass balance simulations using dynamical downscaled data and simulations based on observed temperature and precipitation data are in reasonably good agreement when calibration is used to minimize systematic biases in the MM5 downscaling. These results point to the potential utility of dynamically downscaled future projections derived from high-resolution regional models driven by GCM projections. However, a common drawback is that systematic biases in the GCM projections propagated into the results from regional models.

3.9. Conclusions

This chapter discusses approaches for using the output from current generation climate models for regional climate projection applications. The skill and reliability of individual models will generally vary with the parameter, region, and metric. The reasons for these inconsistencies are rarely clear, which bears on the credibility of the composite model results on regional scales. Nevertheless, there appears to be utility in regional projections, as gauged by the hundreds of studies noted by the Program for Climate Model Diagnosis and Intercomparison (PCMDI).

The strategy for dealing with climate model errors and uncertainties should be keyed in accordance with a user's particular application. First, models should be pre-screened on the basis of their inclusion and formulation of key physical and dynamical processes, especially those directly relevant to the application at hand. Second, the available models can be evaluated based on the ability of their 20th-century hindcasts

to reproduce the large-scale climate variability for the region of interest. This variability can include the annual cycle of certain variables that are responsive to radiative forcing; it can also include the leading spatial and/or temporal modes of variability. Such an evaluation can serve as a basis for eliminating some models from further consideration. Third, hindcasts of specific parameters can be considered in the region of interest. Correspondence of these hindcasts with observations can be the basis of a user's further selection of models or assignments of weights to models. More credence should be given to the projections from the models that are found to replicate past large-scale and externally driven conditions, considering both means and variances and perhaps other criteria such as trends or spatial modes of variability. While it may be tempting to determine a single best model and use its projections, this practice has serious limitations. All climate models are subject to structural uncertainty; the spread in the projections from different models provides a measure of this source of uncertainty.

The Arctic examples presented here suggest that metric-based sub-selection of models is effective for constructing ensemble averages. It is counter-intuitive and counter-productive to include the simulations from models known or strongly suspected to have low skill for a particular regional parameter relative to an observational constraint. While this is a logical consideration, there is no proof that it necessarily provides better projections. Finally, consideration can be given to the sophistication of each model in relation to the parameters of interest. For example, models with superior schemes for handling sea ice are liable to be more appropriate for Arctic applications such as this assessment.

The coarse resolution of most current climate models certainly dictates caution in their application on smaller scales in heterogeneous regions such as along coastlines, sea ice margins, or in locations with rugged orography. The direct use of data from GCMs for many cryospheric applications is not currently feasible, due to the biases in model data on regional scales and due to the inability of models to resolve features of sea ice (e.g., polynyas and leads), glaciers, and topographic features that have major influences on snow and permafrost. In fact, many cryospheric features vary over much smaller scales than climate model grid boxes, and terrestrial snow and ice often occupy complex terrain that is only coarsely resolved in the models' underlying topography.

Even within the reduced set of selected models, the quest for reasons for conflicting model results has no apparent quantitative solution. Given this situation, the use of selection metrics and aggregation methods should remain simple. Moreover, as applications will differ using these practices, choices made for any analysis should be as transparent as possible, with careful documentation of the model selection procedure. Finally, studies that include a selection of models should openly acknowledge that the selection process is guided by the application.

While noting the difficulties in interpreting the reasons for the range of results from different model forecasts for the same variable, overall interpretation of GCM projections has been useful. They include known physics and are quantitative in their calculation of mass and energy budgets. With a careful consideration and selection of models based on past performance compared to observations, the range of results can be reduced, providing a useful prediction of future climate states for regional and larger domains. That Arctic temperatures will generally increase by 3 to 6 °C (the greatest projected increases on the planet), that September will be nearly sea-ice free by mid-century, and that precipitation, but not necessarily as snow, will increase are robust conclusions from considering the results from 23 IPCC GCM models. In this chapter, the high-latitude needs of SWIPA have led to the identification of preferred models for use in the SWIPA modules. The choices of models in the subsequent chapters reflect these findings.

Acknowledgments. The authors acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI), and the World Climate Research Programme's (WCRP's) Working Group on Coupled Modelling (WGCM) for their roles in making available the WCRP CMIP3 multi-model dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy. We appreciate the organizational support to host workshops to discuss the use of CMIP3 models

from the Arctic Council AMAP Climate Expert Group and the SWIPA Integration Team, the Ecosystem Studies of Sub-Arctic Seas (ESSAS), and Working Group 20 (Evaluations of Climate Change Projections) of the North Pacific Marine Science Organization (PICES). James E. Overland, Muyin Wang, and Nicholas Bond appreciate the support of the NOAA Arctic Program Office, the NOAA FATE Project, and the Arctic Yukon Kuskokwim (AYK) Sustainable Salmon Initiative. This chapter is partially funded by the Joint Institute for the Study of the Atmosphere and Ocean (JISAO) under NOAA Cooperative Agreement No. NA17RJ1232, Contribution # 1760. John E. Walsh and William L. Chapman were supported by NSF Grants OPP-0652838 and OPP-0520112. Vladimir M. Kattsov was supported by NSF Grant OPP-0652838 and Russian Foundation for Basic Research Grant 08-05-00569. Pacific Marine Environmental Laboratory contribution #3428.

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Chapter 4: Changing Snow Cover and its Impacts

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Key Findings

- Snow cover changes over the Arctic in response to warming and increasing winter precipitation occur at different rates and in different directions depending on the season, region, and snow cover variable.

- Changes in Arctic snow cover are contributing to changes in the Arctic thermal and hydrological regimes, with important implications for permafrost, Arctic ecology, and snow-vegetation-climate feedbacks.

- Climate model projections of changes in Arctic snow cover over the next 50 years from the models used in the IPCC Fourth Assessment are generally consistent with observed trends, but they have large uncertainties.

- The changing Arctic snow climate is already generating widespread human, ecological, and economic impacts, which will probably intensify in the future. Impacts can be beneficial or deleterious.

- Adapting to changing snow cover conditions is an ongoing process; however, the current rate of change, together with new constraints, will challenge adaptive capacity.

Summary

Snow is a dominant feature of the Arctic terrestrial landscape for eight to ten months of the year. Its extent, dynamics, and properties affect climate, human activities (including socio-economics and culture), infrastructure, hydrological processes, permafrost, hazards, biodiversity, and ecosystem processes. This chapter provides a baseline assessment of recent past and projected future changes in snow cover and their likely consequences.

Observed changes in snow

The ability to monitor changes in Arctic snow cover has improved markedly over the past ten years with an increase in the frequency and resolution of satellite data as the MODIS, AMSR-E, and QuikSCAT satellite systems have provided high-resolution information on snow cover (extent, water equivalent, and melt dates) at the ends of the spectrum. However, major gaps in surface observing (visible and microwave) networks and inconsistencies in re-analysis products pose challenges for determining past variability and the rate of change in Arctic snow cover. Furthermore, the provision of accurate precipitation information over the Arctic remains an ongoing challenge as the observations are prone to significant random errors and biases and automation of observations has created new challenges for maintaining the integrity and quality of observations.

Analysis of *in situ* and satellite snow cover data shows evidence of different regional snow cover responses to the widespread warming and increasing winter precipitation that has characterized the Arctic climate for the past 40 to 50 years. The largest and most rapid decreases in snow water equivalent (SWE) and snow cover duration (SCD) are observed over maritime regions of the Arctic with the highest precipitation amounts (Alaska, northern Scandinavia, and the Pacific coast region of Russia). There is also evidence of marked differences in the response of snow cover between the North American and Eurasian sectors of the Arctic, with the North American sector exhibiting decreases in snow cover and snow depth over the entire period of available *in situ* observations from around 1950, while widespread decreases in snow cover are not apparent over Eurasia until after around 1980. However, snow depths are increasing in many regions of Eurasia. This discrepancy warrants further study as both continents have experienced significant increases in winter precipitation over the past 50 years.

Seasonal differences in observed changes in snow cover are also marked, with most of the decreases in SCD occurring in the spring period, consistent with stronger warming and positive snow- and ice-albedo feedbacks in the spring and the effect of black carbon in decreasing snow albedo (reflectivity of the snow). New analyses of *in situ* observations and satellite data show stronger decreases in snow cover in Arctic coastal regions than inland, likely to be in response to earlier sea-ice disappearance.

Warming and more frequent winter thaws are contributing to changes in snowpack structure, which has important implications for land use and provision of ecosystem services. For example, increased numbers of winter thaws and rain-on-snow events are associated with an increase in potential for ice crust formation, but a shorter melt season has been found to reduce the frequency of basal ice formation over the Russian Arctic west of the Taymir Peninsula. Surface crusts and basal ice layers hinder the ability of reindeer/caribou and other animals to feed.

Projected changes in snow cover

Although climate model projections of changes in Arctic snow cover over the next 50 years from the models used in the Fourth Assessment of the Intergovernmental Panel on Climate Change (IPCC) are for the most part consistent with observed trends, there are large uncertainties in the projections related to inadequacies in the representation of Arctic climate, snow processes, and snow-albedo feedbacks. Further uncertainties arise because important Arctic snow processes such as blowing snow and vegetation interactions are not included, there are large model differences in snow-albedo feedbacks, and models tend to have cold-wet biases over higher latitudes.

The CMIP3 (Coupled Model Intercomparison Project phase 3) general circulation model (GCM) consensus of projected changes in snow cover over the Arctic in response to increasing greenhouse gas concentrations in the atmosphere shows three regions of SWE response: significant decreases in maximum SWE over temperate regions such as Scandinavia; a broad zone over the boreal forest region without statistically significant changes; and a northernmost zone of increased maximum SWE over northern Siberia and the Canadian Arctic Archipelago. In contrast, SCD is projected to decrease over the entire Arctic with the largest and earliest changes over more temperate regions and the smallest and slowest changes over high latitudes. The magnitude of the projected changes by 2050 is in the range of 0 to 15% for maximum SWE over much of the Arctic, with the largest increases (15% to 30%) over the Siberian sector. Annual SCD is projected to decrease by about 10% to 20% over much of the Arctic, with the smallest decreases over Siberia (<10%) and the largest decreases over Alaska and northern Scandinavia (30% to 40%).

Impacts

Changes in precipitation and temperature are expected to dramatically change the hydrological regime of the Arctic by affecting snow accumulation and melt, evaporation, and runoff as well as short- and long-

term water storage changes. However, the net hydrological balance for ecosystems and biogeochemical cycling is uncertain.

Arctic ecological processes interact with snow cover in many complex ways. Vegetation affects interception, sublimation, snow trapping, energy exchanges, hydrology, and accumulation and ablation of snow. Snow provides protection for vegetation from extreme environments and provides water in the spring. Interactions between snow and current and projected advances of shrubs and trees into the tundra are expected to amplify warming in the Arctic. Changing snow conditions also affect biogeochemical cycles. The date of the onset of snow melt is a good predictor of annual atmospheric carbon accumulation as the timing and duration of the snow-free season control the growing season for vegetation. Geochemical processes in the snowpack affect the concentrations of trace gases in the Arctic's lower atmosphere. Deposition of contaminants to snowpacks concentrates many compounds, some of which can be hazardous as they move through the food chain. A shorter snow period would reduce the period for accumulation.

Socio-economic consequences

Projected changes in the hydrological cycle will have impacts on several commercial sectors. A more even water discharge, resulting from changes in snow cover and duration, will affect the capacity and operations of current and future hydroelectric developments and might solve some of the rising energy needs. A more even distribution of water discharge will also reduce the wastage of peak reservoir levels that do not generate maximum electricity. Despite reduction in snow cover in the Arctic, the northward retreat of snow cover in mid-latitudes may provide new opportunities for snow-based tourist activities in the Arctic. However, changes in the consistency of the snowpack and the unpredictability of snow conditions could lead to negative experiences for tourists and even an increased frequency of hazards such as snow avalanches and slush torrents. Furthermore, reduced Arctic snow cover is already limiting the productivity of some forests. Although boreal forestry is expected to become more productive and move northward during climate warming, snow is an important source of water that might become limiting. As snow cover is reduced and rainfall increases, broadleaf trees will have more advantage than conifers, thereby decreasing carbon storage and negatively impacting the forest economy. Changing snow conditions have damaged valuable timber species leading to the loss of millions of hectares of trees.

Traditional land use is tightly coupled to snow conditions. Snow depth, distribution within the landscape, timing of snow onset and thaw, extreme climate events in winter, and the physical properties of snow throughout the profile of the snowpack all have impacts on reindeer/caribou and their management as well as general activities (such as transport) of Arctic residents.

Effects of climate change on human health include direct impacts (such as those caused by changes in temperature, snow cover, and ultraviolet light) and indirect impacts (such as those caused by climate-induced changes in wildlife and the diseases that they share with humans). The impact of changing snow on subsistence wildlife species is critical to the diet of indigenous residents, and its impact on recreation and mobility could affect psychosocial stress. Already, many Arctic indigenous people and other residents have high and increasing mortality rates from work-related accidents sustained while undertaking hazardous tasks in harsh environments.

Adaptation and mitigation

Arctic communities have limited ability to mitigate likely future changes in snow, and they must adapt to transitional snow regimes. However, they cannot migrate as freely as during historic climatic changes. Adaptation will be necessary in the areas of infrastructure maintenance and development, land use and resource management, reindeer herding, conservation, tourism, hunting and fishing, agriculture, and forestry. Flexibility, diversification, and investment will be required to ensure successful adaptation. However, despite their ability to adapt, Arctic peoples' subsistence livelihoods are likely to become more

insecure, and there is likely to be a shift from a mixed economy to a market-based economy that will have significant cultural implications.

Improved monitoring and forecasting of snow conditions and a better understanding of relevant projections of snow are needed at the local scale for adaptation. The provision of relevant information from scientists requires the active involvement of community members; local, regional, and national organizations; and decision- and policy-makers. Greater engagement is also needed between the scientific community and Arctic residents in order to incorporate traditional knowledge in the design and interpretation of monitoring and modeling studies.

The multitude of impacts responding to changing snow conditions provides challenges and opportunities. The ability to adapt to the challenges and respond to the opportunities will vary between urban and rural populations and the Arctic and global communities: there will be winners and losers. Policy development is required to enhance the adaptation potential and opportunities and to minimize the challenges for the potential losers.

4.1. Introduction

Frozen precipitation accumulating on a surface creates a snow cover. Snow is an important and dominant feature of Arctic terrestrial landscapes with cover present for eight to ten months of the year. Its extent, dynamics, and properties (e.g., depth, density, water equivalent, grain size, and changes in structure throughout its vertical profile) affect climate (e.g., ground thermal regime), human activities (e.g., transportation, resource extraction, water supply, use of land, and ecosystem services), as well as infrastructure, hydrological processes, permafrost, extreme events (including hazards such as avalanches and floods), biodiversity, and ecosystem processes. Snow is therefore a significant component in the socio-economics of Arctic societies. Snow also has a number of important linkages to other components of the cryosphere (see Chapter 1). Large-scale anomalies in snow cover extent, depth, and water equivalent tend to persist for longer (for months or even for most of the cold season) than the synoptic conditions through which they were initially caused. Therefore, snow cover at monthly to seasonal timescales becomes one of the few significant memory components of the Earth's climate system and is used in long-term weather forecasting, which has many profound implications for people. Over millennia, snow has provided an archive of past environments and climates that is currently being used to investigate the causes of climate change (see also Chapter 8 on the Greenland Ice Sheet).

Air temperature and precipitation are the main drivers of regional-scale snow cover variability over the Arctic region, with local-scale variability in snow cover related to interactions with vegetation cover and terrain through processes such as blowing snow, canopy interception, and sublimation (when water changes directly from solid to vapor form without thawing) (King et al., 2008). Impurities in the snowpack (e.g., leaf litter and organic and black carbon) contribute to local (landscape) and regional (circum-Arctic) variations in surface albedo, which influence the surface energy budget and spring season melt rates. Over Eurasia, the climate-forcing effect of surface deposition of organic and black carbon in reducing snow surface albedo is estimated to be of the same order of magnitude as that of anthropogenic carbon dioxide (Flanner et al., 2008; Grenfell et al., 2009).

Observational data show that the Arctic's land area has experienced a significant poleward amplification of global warming trends and increased precipitation over the past several decades (Déry and Brown, 2007; Trenberth et al., 2007; Min et al., 2008). Consequently, although there is some evidence of increasing snow accumulation over northern Eurasia (Kitaev et al., 2005; Bulygina et al., 2009), the dominant trend in recent decadal snow cover is earlier spring snow melt and springtime declines in snow cover associated with more frequent rain-on-snow and mid-winter melt events in some regions of the Arctic (Groisman et al., 2003; Burn et al., 2004).

These trends in the main drivers of snow variability are likely to continue. According to the IPCC Fourth Assessment (IPCC AR4) (Christensen et al., 2007), the Arctic is *very likely* to warm during the 21st century, particularly during winter. Annual Arctic winter precipitation is also considered *very likely* to increase. This degree of certainty in projected precipitation changes is much higher over the Arctic than other regions of the globe as there is a high level of consistency between climate model projections of precipitation changes in this region. The CMIP3 suite of coupled climate models (Meehl et al., 2007) projects reduced SCD over the Arctic with increased annual maximum snow accumulation over the northern extremities of both continents (Lemke et al., 2007; Räisänen, 2007; Brown and Mote, 2009). However, the rate and magnitude of the snow cover response to changing climatic conditions is expected to vary considerably due to regional differences in projected climate change and differing regional and elevational sensitivities of snow to climate change (Brown and Mote, 2009). The response of snowpack properties to a changing climate is also potentially complex to project, because in addition to affecting start and end dates of snow cover, changes in air temperature affect the fraction of precipitation falling as snow, rain-on-snow events, freeze-thaw events, the density of snowfall, and the density of snow on the ground. Increasing winter precipitation may result in increased accumulation in some areas, while enhanced growth of Arctic shrubs in response to warming (e.g., Sturm et al., 2001b; Forbes et al., 2010) may result in local increases in snow cover accumulation independent of increases in precipitation.

Changing snow cover in the Arctic has many complex implications for the natural world and for the economic development of society, including indigenous peoples and other Arctic residents, through changes in the physical environment and the associated impacts on climate feedbacks, water resources, infrastructure, and ecosystem services. For example, reductions in snow cover extent and duration as well as decreased reflectance from accumulation of black carbon and dynamic vegetation interactions provide positive feedbacks to the climate system in that they enhance the rate of warming over the Arctic (Hansen and Nazarenko, 2004). In addition, it is anticipated that future variations in snow depth may dramatically change the hydrological regime of the Arctic (Adam et al., 2009), which in turn will affect water resources, transportation, and ecosystem services. Changes in snow accumulation will also affect the ground thermal regime with impacts on permafrost temperature and distribution. Changes in snow regime and their effects on permafrost affect hydrology, infrastructure (buildings, runways, pipelines, transport routes), and ecosystems (see Chapter 5). Changes in snowfall regime (amount, timing, and phases of water or snow) over the Greenland Ice Sheet and other Arctic glaciers and ice sheets will affect the mass balance of the ice (see Chapters 7 and 8), while more frequent mixed-precipitation events and rain-on-snow events will have adverse impacts on biodiversity, animal populations, vegetation, and ecosystem processes. Changes in snow accumulation also affect freshwater availability and sea-ice growth rates, light transmission through ice, and ice dynamics (see Chapter 9). Deposition of airborne contaminants on snowpacks concentrates many chemical compounds. Snowpack chemical reactions affect nitrogen oxides, halogens, ozone, organic compounds, and mercury, which affect people and terrestrial, freshwater, and marine Arctic ecosystems (AMAP, 1998, 2009).

The identification of snow-related vulnerabilities and sensitivities, together with some assessment of the uncertainties involved, is essential to determine the information required for developing adaptation responses and policies. Arctic societies have constantly adapted to changing climatic conditions, but their increasing reliance on urban environments and infrastructure now limits some of their options (Chapin et al., 2004). However, current ability to predict future changes in climate provides advanced warning of impacts and opportunities that will allow early development of adaptation strategies. In addition, development of climate downscaling techniques provides some of the tools required for adaptation planning.

Although changes in snow cover and its properties are already significant and future changes will have many socio-economic and other implications, snow has not been treated authoritatively in previous assessments. For example, the Arctic Climate Impact Assessment (ACIA, 2005), and the response of the Arctic snow cover to climate change simulated by the CMIP3 models was not addressed in the IPCC AR4

(Solomon et al., 2007). For these reasons, this chapter cannot use the ACIA report or the IPCC AR4 as benchmarks, and so draws on earlier, as well as more recent work, to provide a new benchmark.

This chapter aims to provide a synthesis of current knowledge of existing and projected changes in Arctic snow cover (including various properties of snow) and the likely impacts on society. The chapter presents basic information on snow characteristics and their relevance to the natural world and to society, charts recent changes in snow conditions, and synthesize new projections of changes in snow and their implications. The information is used to highlight how adaptations to changing snow conditions might occur in the natural world and how society might develop socio-economic adaptation strategies. Data are drawn from many different sources, including monitoring programs, field experiments, model output, and traditional knowledge of indigenous peoples. There is a particular focus on knowledge from new sources such as International Polar Year (IPY) projects. The chapter ends by listing important gaps in understanding and recommends priority actions.

4.2. Basic snow characteristics from an Arctic perspective

- Snow is a dominant feature of the Arctic landscape persisting for eight to ten months of the year.
- Snow depth, extent, duration, timing, water equivalent, and stratigraphy have many consequences for the climate system, hydrology, permafrost, ecology, biogeochemical cycling, and socio-economics.

4.2.1. Characteristics and definitions of Arctic snow

Frozen precipitation accumulating on a surface creates a snow cover. In the Arctic, snow covers land and ice surfaces for eight to ten months of the year, with important regional gradients in duration and amount related to topography and regional variations in temperature and precipitation. A summary of the key features of the Arctic snow cover is provided in Box 4.1 (see Sturm et al., 1995 and Fierz et al., 2009 for scientific terminologies).

Box 4.1. Arctic snow cover

The characteristics of Arctic snow cover are the result of a complex interplay of atmospheric and surface processes that determine not only the quantity of water stored as snow, but also snowpack condition (e.g., grain size, density, ice layers). The amount of snow accumulating on a surface is influenced by precipitation amount, type, and timing; blowing snow transport and sublimation; and vegetation interception. However, the character and evolution of high-latitude snowpack has the additional complexity of being particularly strongly dependent on blowing snow processes with the distribution and physical properties of snow on the ground closely linked to local-scale variability in terrain and vegetation (King et al., 2008).

The key large-scale physiographic and climatic factors influencing the regional distribution of Arctic snow cover (see Figure 4.1) are elevation, amount of vegetation cover, spatial distribution of freezing temperatures, and location of the main cyclone tracks bringing moisture into the Arctic. Air temperature and elevation exert the strongest influences on the distribution of SCD across the Arctic (Figure 4.1e) with both continents exhibiting marked east-west increases in snow cover in response to the modification of winter air masses over the cold, dry continental interiors. Land areas in the zone of -20 °C temperatures (see darker blue area in Figure 4.1c) experience snow cover for most of the year, and climate models suggest that this area will see increased snow accumulation in the future (Räisänen, 2007). The spatial distribution of SWE is more complex than SCD but is basically driven by moisture availability over the snow season, reflected in the cyclone frequency map (Figure 4.1d). The highest snow accumulations in the Arctic are located in the coastal mountain regions and considerably more moisture penetrates into the western sector of the Eurasian Arctic than North America, where the coastal mountains block moisture entering from the Pacific Ocean. Regions with winter temperatures closer to freezing, such as Scandinavia and the Pacific coasts of Russia and Alaska, are also more likely to experience thaw and rain-on-snow

events that create ice layers in the snowpack. Snow cover in these maritime regions of the Arctic has been shown to be more sensitive to temperature changes (Brown and Mote, 2009).

The high winds, low temperatures and low snowfall amounts over the exposed tundra regions of the Arctic produce a snow cover that is typically quite shallow, about 30 to 40 cm (except in drifts and gullies), with a wind-hardened surface layer ('wind slab') overlying a less dense depth hoar ('sugar snow') layer (Derksen et al., 2010). The average snow density remains close to 300 kg/m^3 over much of the snow season, but snow depth and properties can exhibit strong local variation with many exposed areas, drifts, dunes, and sastrugi (sharp irregular ridges on the snow surface formed by wind erosion and deposition). In forested regions of the Arctic (taiga and boreal forest), snow cover is more uniform and less dense (~200 kg/m^3) as the trees act as windbreaks and shade the snow from incoming solar radiation in the spring (McKay and Gray, 1981). In contrast, north of the treeline, where wind action compacts the snow, snow density is higher.

The average seasonal evolution of snow depth and density at three Canadian Arctic sites in different climate regimes – Fort Reliance in the sub-humid high boreal zone, Baker Lake in the low Arctic zone, and Clyde in the high Arctic zone (zones as defined by Environment Canada, 1989) – show characteristic linear increases in SWE and snow density over the snow season typical of Arctic snow cover (Figure 4.2). Clyde has the largest accumulations of the three sites due to its location on the east coast of Baffin Island where winter cyclones are more frequent (Figure 4.1d).

Snow is critically important to the lifestyle and well-being of indigenous peoples. The Inuit have a rich snow vocabulary with many of the words related to uses of snow for drinking water, making shelters, and trafficability (see 'snow' in the online English-Inuktitut-French glossary maintained by Nunavut Arctic College at www.btb.gc.ca/btb.php?lang=eng&cont=202#a). Similarly, the Sámi have a rich vocabulary, but this also relates to snowpack conditions relevant to reindeer herding practices (Ryd, 2001).

There is a wide range of regularly observed snow cover information in the Arctic from *in situ* and satellite observations (see Appendix 4.1 for a detailed discussion). The SCD on the ground is one of the best-observed variables in terms of resolution and longevity. Snow depth and SWE (the depth of liquid water that would result from melting the snow) are more difficult to monitor due to their high spatial variability, large gaps in the *in situ* observing networks, and difficulties in monitoring from satellites. Information on snow density and surface properties is important for transportation and reindeer herding in the Arctic; however, these properties are routinely measured at only a limited number of research stations outside the Russian Federation, where they form part of the snow course observation routine at the national meteorological network, in operation since 1966 (Bulygina et al., 2010a). In some remote areas, for example in Zackenberg (northeast Greenland), a digital camera placed in a weatherproof box with battery and solar cells has been used to record snow depth on stakes during winter and then collected during summer when the site was visited (Hinkler et al., 2002). Long-term records of snow characteristics in the Arctic are rare, with the exception of the record since 1913 from Abisko in the Swedish sub-Arctic (Kohler et al., 2006).

The indigenous peoples of the Arctic have a profound knowledge of changing snow conditions of practical importance for survival, which has been passed from generation to generation. Languages, such as the Sámi language and Inuktitut, have many terms to describe snow cover related to migrations and transportation, hunting, and reindeer herding (Ryd, 2001). The snow surface is often described from a mobility perspective, while the descriptions of snow layers detail animals' ability to find grazing. Such knowledge has created and preserved the way of life for indigenous peoples in the Arctic (e.g., Ryd, 2001; Roturier and Roue, 2009) and this depth of understanding cannot be fully captured in standard weather and hydro-meteorological monitoring programs.

4.2.2. Role of snow cover in the Arctic climate system

Snow cover has a number of important physical properties that exert an influence on climate or moderate its effects (see Cohen and Rind, 1991). It has high short-wave albedo, high thermal emissivity, low heat conductivity, large latent heat of fusion, and low surface roughness while it stores and rapidly releases water in the melt season. The combination of high albedo (as high as 0.8 to 0.9 for dry snow) and low thermal conductivity promotes low surface temperatures and low-level temperature inversions. The low thermal conductivity of snow allows it to insulate the surface from large energy losses in winter, and this has major implications for the energy and moisture fluxes in the near-surface layers. Consequently, the insulating effect has a strong impact on ice growth rates and ice thickness and on the development of seasonally frozen ground and permafrost. The surface energy exchanges are further modified by the low aerodynamic roughness of snow cover, which can reduce turbulence and, hence, vertical transfer of energy. On some occasions, sastrugi and wind scoops may negate the streamlined surface by interaction with turbulence of strong winds and may act as a feedback loop to enhance turbulence over snow cover. However, the major snow impact on surface roughness is its reduction.

Onset of snow cover is associated with an abrupt drop in surface air temperature (by up to 10 °C; Voeikov, 1952; Leathers and Robinson, 1993; Groisman et al., 1997). This is mostly caused by a dramatic decrease in the surface energy budget due to the high reflectivity of newly established snow cover that enhances and stabilizes the temperature decrease. However, about half of the temperature decrease is caused by the weather factors that have already decreased air temperature (i.e., the cold fronts that brought snowfall in the first place) (Voeikov, 1952; Lamb, 1955; Leathers et al., 1995). In the mid-latitudes, it is usually a few weeks between the first snowfall and the formation of a continuous snow cover, but in the Arctic one major snowstorm in autumn may be enough to generate a continuous snow cover, because there is insufficient energy at this time to reverse the state of the ground to snow-free conditions. For example, at 60° N in October over Eurasia the surface radiation budget is in the range of ± 10 W/m^2 (see Figure 17 in Groisman and Bartalev and NSPD Team, 2007), which is insufficient for appreciable snow melt. Thus, once established, snow cover feedbacks tend to support the continuing existence of the snow cover. A number of studies have shown that large-scale snow cover anomalies can play a significant role in global-scale atmospheric circulations that act over timescales of seasons to years (Groisman et al., 1994a; Gong et al., 2004; Fletcher et al., 2009a,b).

When snow cover is established, its insulating properties affect both the surface and the near-surface temperature regime. Soil surface temperature in mid-winter under 50 cm of snow may be around zero, while the air temperature above the snowpack is as low as -20 °C (Pomeroy and Brun, 2001). The insulating role of snow prevents deep freezing of near-surface ground, which has important implications for the active layer and for permafrost (see Chapter 5), as well as for the runoff response in spring and for ecological and biogeochemical processes. In the permafrost areas of the Arctic, variations in snow cover depth (rather than surface air temperature) are a major factor controlling temperature variability in the upper 3 m of soil. Due to an increasing winter snow depth in these regions, soil warming trends are observed even at locations with cooling trends in surface air temperature (Bulygina et al., 2007, 2009). At these sites, the contribution of the mean annual air temperature to the total variance of soil temperature at 160 cm is less than 5%, while the contribution of the snow depth is 30% (Sherstyukov et al., 2008).

During spring, when there are large amounts of incoming solar radiation (greater than 100 W/m^2), the presence of snow on the ground effectively delays warming. This is because of the still relatively high reflectivity of melting snow (albedo ~0.5 to 0.7) compared to other surfaces, such as soil, vegetation, or ponds on the ground or on sea ice, and the energy required to melt it. The large amount of energy required to melt snow means that near-surface spring temperatures stay close to 0 °C with frequent melt-and-refreeze cycles until the snow cover becomes discontinuous. These freeze-thaw cycles are readily observable with satellite data (Rawlins et al., 2005; Bartsch et al., 2007). Once the snow cover becomes discontinuous, the melting process becomes rapid owing to advection of energy from snow-free patches (Marsh et al., 1997). Nevertheless, even with high levels of insolation it may take more than a month for the snowpack to disappear completely in regions with a deep snow cover, in shaded areas, or where deep drifts have formed (Ramage et al., 2007).

Changes in snow cover over the Arctic region, particularly in spring, have a strong impact on regional energy budgets owing to the large amount of incoming solar energy reaching the snow surface (Groisman et al., 1994a,b; Déry and Brown, 2007). From a modeling study, Euskirchen et al. (2007) estimated that a pan-Arctic reduction in snow cover of 0.22 d/y over the 1970 to 2000 period of warming contributed an additional 0.8 W/m^2 per decade of energy. Their simulation also showed evidence of regional asymmetry in Arctic snow-albedo feedback with stronger heating contributions over the North American sector from loss of snow cover. Recent work by Fernandes et al. (2009) highlighted the importance of snow metamorphism (or aging) in the albedo feedback with their satellite-based analysis showing that snow metamorphism was as important as loss of snow cover in determining the total terrestrial snow-albedo feedback over Eurasia.

4.3. Changes in snow regime and properties

- Snow cover duration has decreased. The largest and most consistent change in snow cover is earlier disappearance of snow in spring, averaging 3.4 days per decade over the pan-Arctic terrestrial region (excluding Greenland) during 1972 to 2009. The larger response in spring compared with autumn (0.5 days per decade) is consistent with greater positive snow-albedo and black carbon feedbacks in spring.
- Snow cover extent has decreased. Data from visible satellite imagery show that Arctic snow cover extent in May and June decreased by an average of 18% over the 1966 to 2008 period of the National Oceanic and Atmospheric Administration (NOAA) record.
- Rates of change in snow water equivalent and snow cover duration vary across the Arctic. The largest decreases in snow water equivalent and snow cover duration occur in maritime regions of the Arctic (Alaska, northern Scandinavia, and the Pacific coast region of Russia), while new analyses show stronger decreases in Arctic coastal regions than inland.
- Snow depth changes have differed between North America and Eurasia over the past 50 years. Many regions of Eurasia exhibit increasing maximum *in situ* snow depth trends, while North American sites show decreasing trends, although both continents have experienced significant increases in winter precipitation over the past 50 years.
- There is new evidence of changes in snowpack structure, such as more frequent ice crust formation resulting from more frequent winter thaws and rain-on-snow events. These changes have important implications for land use and provision of ecosystem services.
- Annual snow cover duration is projected to decrease. Decreases by 10% to 20% over most of the Arctic by 2050 are projected, with the smallest decreases over Siberia (<10%) and the greatest losses over Alaska and northern Scandinavia (30% to 40%). The earliest and largest decreases in snow cover duration and accumulation are projected to occur over coastal regions of the continents in agreement with observed trends.
- Slight increases in maximum snow accumulation are projected. Increases of 0 to 15% are projected over much of the Arctic with the largest increases (15% to 30%) over the Siberian sector.
- The frequency and areal extent of rain-on-snow events are projected to increase over all regions of the Arctic over the next 50 years.

4.3.1. Observed changes in Arctic snow cover and snowfall

4.3.1.1. Changes in solid precipitation

In the Atlantic, North European, and West Siberian sectors (see Appendix 4.1 for sector definitions), the climatic conditions are formed largely under the influence of heat and moisture advection from the North Atlantic area. Climate in the East Siberian and Chukchi sectors is significantly influenced by circumpolar conditions over the northern Pacific Ocean, as well as by the center of action above Siberia. The Alaskan sector is also influenced by circumpolar processes over the northern Pacific Ocean. In the Canadian sector, the climatic conditions in winter are governed both by anticyclonic circulation above northwestern Canada and the Arctic Basin and by a frequent passage of Atlantic lows.

The frequency, duration, and intensity of snowfall play a major role in the formation, characteristics, and variability of a snow cover. Analysis of trends in seasonal totals of precipitation from October to May

(which correspond to the snowfall season with mean monthly temperatures below -2 °C) at climate stations located north of 60° N revealed an increase in cold season precipitation between 1936 and 2009 in almost all sectors of the Arctic (Figures 4.3 and 4.4; Table 4.1).

Table 4.1. Linear trend analysis (least-squares method) for cold season (October to May) total precipitation over the long term (1936 to 2009) and short term (1980 to 2009). See Appendix 4.1 for sector definitions. Values in **bold** are significant at the confidence level $p \leq 0.10$ (B_x = trend in mm per decade; Δ = linear change over the period as a percentage of the 1961 to 1990 mean; r^2 = percentage of variance explained by linear regression.

Areas	60 – 70 ° N						70 – 85 ° N						60 – 85 ° N					
	1936 – 2009			1980 – 2009			1936 – 2009			1980 – 2009			1936 – 2009			1980 – 2009		
	B_x	Δ, %	r^2, %	B_x	Δ, %	r^2,%	B_x	Δ, %	r^2,%	B_x	Δ, %	r^2,%	B_x	Δ, %	r^2,%	B_x	Δ, %	r^2,%
Atlantic (50° W to 30° E)	9.2	8.2	22.2	-13.7	-4.8	13.0	**9.1**	28.6	46.7	10.1	12.9	27.1	**9.1**	10.3	30.8	-5.5	-2.6	7.4
North European (30° E to 60° E)	**16.2**	40.0	70.8	7.9	7.8	19.0	**6.2**	20.3	50.9	9.9	13.1	29.8	**12.1**	33.0	73.8	**9.0**	9.9	30.9
West Siberian (60° E to 100° E)	**10.3**	28.5	65.4	1.2	1.3	4.5	-0.2	-1.1	1.9	**17.1**	38.6	65.2	**5.9**	20.6	59.0	**7.9**	11.2	37.3
East Siberian (100° E to 150° E)	**3.1**	15.0	41.3	4.5	8.8	26.4	**-3.5**	-21.2	38.8	1.0	2.5	5.2	0.6	2.7	9.7	3.1	6.7	22.3
Chukchi (150° E to 170° W)	-0.2	-0.7	1.3	-6.9	-9.5	24.8	**-8.4**	-42.3	56.4	-3.3	-6.7	17.3	**-3.4**	-12.8	24.3	-5.5	-8.7	29.3
Alaskan (170° W to140° W)	2.5	10.9	15.1	-0.0	0.0	0.0	-0.4	-6.6	3.6	**10.1**	67.8	44.1	2.0	9.4	14.7	1.7	3.4	5.4
Canadian [a] (140° W to 60° W)	**6.3**	23.1	26.9	8.1	11.7	8.1	**8.5**	54.2	28.7	2.9	7.6	0.8	**7.8**	30.7	40.3	**10.0**	16.5	17.0
Whole latitudinal zone [b]	**5.7**	14.5	27.2	0.1	0.1	0.0	**1.8**	10.5	7.4	**5.8**	13.7	12.1	**3.6**	10.9	20.6	3.7	4.7	5.2

[a] For Canada the adjusted and homogenized precipitation dataset of Mekis and Hogg (1999, updated to 2008) was used. Data are available from 1936 to 2008 for 60° to 70° N, but there are no data prior to 1939 for 70° to 85° N (and 60° to 85° N); [b] for non-overlapping sectors weighted by land area for 1936 to 2008 for 60° to 70° N and 1939 to 2008 for 70° to 85° N and 60° to 85° N.

The analysis is based on monthly total precipitation data collected at the stations from the start of their operation up to 2009. Regionally averaged series were calculated over the latitudinal zones from 60° to 70° N, 70° to 85° N, and 60° to 85° N using a method of optimal averaging for the seven sectors shown in Table 4.1 (Frolov, 2009). Canadian results were computed from a 2008 update of the adjusted and homogenized precipitation dataset of Mekis and Hogg (1999), but there were insufficient station data in the Canadian Arctic sector north of 70° N to compute trends in the 70° to 85° N and 60° to 85° N sectors prior to 1939 (Appendix 4.1 Figure A2).

4.3.1.2. Snow depth, snow water equivalent, snow cover duration and extent

There have been long-term increases in winter snow depth over northern Scandinavia and Eurasia (Figures 4.5 and 4.6) but significant decreases over the North American Arctic between 1950 and 2006 (Figure 4.7) (Kohler et al., 2006; Bulygina et al., 2009). Similarly, SWE increased over Eurasia and most of northern Russia for the 1966 to 2009 period (Bulygina et al., 2010b), but decreased over northern Canada over the 1966 to 1996 period (Atkinson et al., 2006). This contrast is surprising as both continents have experienced long-term increases in cold season precipitation (Table 4.1, Figures 4.3 and 4.4; Trenberth et al., 2007; Min et al., 2008).

Over most of Russia, the number of days with snow depth greater than 20 cm has increased (Figure 4.8). In the northern and southern regions of Western Siberia, in Yakutia, and on the coast of the Sea of Okhotsk, the trend is 8 to 10 days per decade. In contrast, the Chukchi Peninsula and Transbaikalia has experienced a decrease of days with snow depth greater than 20 cm by 6 to 10 days per decade (Bulygina et al., 2009).

Trends in SCD also contrast between the two Arctic regions (Figures 4.6 and 4.7) but began to decline over Eurasia after 1980. Maximum decreases in SCD of the order of four to six days per decade were found in the western and southern regions of European Russia from 1951 to 2006 (Kitaev et al., 2006; Razuvaev and Bulygina, 2006; Bulygina et al., 2009, 2010b; Shmakin, 2010). In contrast, there is an increase in SCD recorded in Yakutia (central Siberia) and in some regions of the Russian Far East, owing to their proximity to the Pacific.

Groisman et al. (2006) reported a general increase in SCD over Russia and the Russian polar region north of the Arctic Circle by five days (3%) and twelve days (5%), respectively, over the period 1936 to 2004. Linear trend analyses applied to the entire period of observations available for their study lead to the paradoxical conclusion that the increase in SCD could not be associated with 'Arctic warming', which was not apparent over this particular time period.

In the coastal zone of the Eurasian Arctic, the interannual variability of dates of formation and decay of continuous snow cover is large, varying by 1.5 months. The duration of the period with continuous snow cover varies from 200 to 300 days at different stations in this zone (Radionov et al., 2004a). In the Fennoscandian and Alaskan-Canadian sector there has been a statistically significant decrease of about 3 days per decade in SCD (Table 4.2). In contrast, an increase by 1.5 days per decade was detected in the Kara Sea sector. During the past three decades (1978 – 2007), there has been a statistically significant decrease in SCD in coastal and island areas in all sectors of the Arctic (ranging from 4 to 9 days per decade) except for the Kara Sea and the Chukchi Sea sectors (Table 4.2).

Table 4.2. Results of linear trend analysis (using least-squares method) of continuous snow cover duration[a] from regionally averaged *in situ* observations in coastal and island areas of the Arctic from 1951 to 2007. Values in **bold** are significant at the 0.05 significance level. (B_x = coefficient of linear trend, number of days per decade; Δ = linear change over period as a percentage of the 1961 to 1990 mean; r^2 = percentage of variance explained by linear trend). Source: updated from Radionov et al. (2004a).

	1951 – 2007			1978 – 2007		
	B_x	Δ %	r^2 %	B_x	Δ %	r^2 %
Fennoscandia sector	**-3.45**	-10.3	48.0	**-7.34**	-11.5	54.1
Barents Sea sector	-0.59	-1.7	9.2	**-6.28**	-9.4	50.1
Kara Sea sector	**1.46**	3.2	29.5	0.98	1.1	11.3
Laptev Sea	-0.26	-0.6	6.5	**-5.35**	-6.0	68.3
East Siberian Sea sector	1.77	3.8	22.3	**-8.74**	-10.0	60.6
Chukchi Sea sector	0.27	0.6	4.9	-2.96	-3.7	30.2
Alaska and Canada sector [b]	**-3.00**	-6.7	33.2	**-4.06**	-4.8	29.0

[a] Defined by Radionov et al. (1996) as the duration of the period with continuous snow cover (50% or more of the visible area is reported as 'snow covered'); [b] computed from Canadian and Alaskan daily snow depth data with snow cover duration defined as the number of days in the snow season with daily snow depths ≥ 2 cm following Brown and Goodison (1996).

The NOAA satellite record shows that variability in SCD in the North American and Eurasian Arctic has been more or less in phase since observations began in 1966, with contrasting seasonal patterns of little change in autumn SCD (i.e., the snow cover onset date) while spring SCD (i.e., the snow-off date) underwent a rapid decrease during the 1980s (Figure 4.9). Foster et al. (2008a) attributed this step change to a regime change in the Arctic Oscillation (AO) to more positive values. However, a new multi-dataset study of spring snow cover changes over the Arctic (Brown et al., 2010) suggested that the reduction in spring snow cover is more linear over time and has a closer link to Arctic temperature trends than to the Arctic Oscillation. The asymmetric seasonal response of SCD is consistent with observed warming trends that are likely being enhanced by positive snow-albedo feedbacks (Groisman et al., 1994a; Déry and Brown, 2007).

The trend (days per decade) in the dates of the onset and disappearance of continuous snow cover over the coastal region of northern Russia for the 1973 to 2003 period shows great variability but a slight tendency for later formation of snow cover in autumn and for an earlier reduction of snow cover in spring (Figure 4.10). Negative values indicate a later date of formation or decay of snow cover in comparison with the long-term average date (Radionov et al., 2004a).

Data from visible satellite imagery show that Arctic snow cover extent in May to June decreased by an average of 18% over the 1966 to 2008 period of the NOAA record. Spatial analysis of snow cover trends in the NOAA dataset (Figure 4.11) confirms a number of the previously mentioned characteristics. Specifically, that decreases are more marked in the snow cover melt period than the snow cover onset period, that the date of snow cover onset has become earlier over northern Russia, and that spring decreases over Eurasia tend to be stronger in northern coastal regions. The average change in SCD over the pan-Arctic region (excluding Greenland) was -0.49 days per decade in the snow cover onset period, and -3.43 days per decade in the spring snow cover melt period (Figure 4.11).

Arctic indigenous peoples have also observed long-term changes in snow conditions as these have direct impacts on their livelihoods (Forbes and Stammler, 2009; Bartsch et al., 2010). However, their observations are at the local scale and few are published. Also, most of their observations relate to changes in snow structure and their impacts on the mobility of reindeer and reindeer herders as well as access to vegetation. Observations on snow cover by the Sámi from northern Sweden highlight recent increases in ice crusts and in some areas snow accumulation, with statements such as "*all valleys were snow free during (reindeer) calving in the 1930s*", "*terrain elements that determined animal movements in the summer are now snow covered: reindeer now find new passes and roam over a wider area*", "*snow-covered areas and snow patches persist longer into the summer in high mountain areas*", and "*rapid thaws created problems when moving to summer grazing areas in 1938 – 40"* (Riseth et al., 2010). Many of these observations match or add to nearby climate station monitoring, particularly for the period up to the 1990s (Callaghan et al., 2010).

Arctic snow cover exhibits large interannual variability, linked to large-scale variation in atmospheric circulation, around the previously documented trends. For example, the increase in snow depth over most of northern Eurasia in recent decades can be explained in two ways. First, there was a dramatic retreat in Arctic sea ice at the end of the warm season (Serreze et al., 2007) that left large ice-free or thin-ice areas at the beginning of the cold season in the Eurasian sector of the Arctic Ocean. These allowed additional water vapor influx into the dry Arctic atmosphere that ended up as snowfall further south. Second, more intensive cyclonic circulation and more frequent cyclones (Popova, 2004) related to changes in atmospheric circulation also caused increased snow depth over most of northern Eurasia in recent decades. This circulation change is linked to significant increases in the North Atlantic Oscillation (NAO) index since the 1970s (Popova, 2007). Snow accumulation in Eurasia is also strongly modulated by the Scandinavian pattern (SCAND) that reflects the intensity of blocking anticyclones in Eastern Europe. Snow accumulation is negatively correlated with the SCAND index, and in the 1951 to 1975 period, was the most important circulation pattern influencing variation in snow depth across Eurasia (Popova, 2007). The Pacific North America (PNA) and Pacific Decadal Oscillation (PDO) have been shown to influence climate and snow cover over the western Canadian Arctic (Derksen et al., 2008b) with positive (negative) modes of the PNA and PDO associated with reduced (increased) snow accumulation and a shorter (longer) snow cover season.

4.3.1.3. Changes in snow properties

4.3.1.3.1. Snow structure and snow cover stratigraphy

Data on trends in snow cover stratigraphy are rare (but see Bulygina et al., 2010a, for data relating to Russia; Riseth et al., 2010, for data relating to Sweden; and Gerland et al., 1999, for data related to Svalbard). Nevertheless, they are important for example in determining animal access to food beneath snow, while the presence of weak layers (depth hoar) or slip planes (crusts or ice layers) in the snowpack increases the potential for avalanche release. Sámi traditional knowledge from the Abisko area of sub-Arctic Sweden reports several local changes in the physical properties of the snow cover, particularly the development since the 1980s of more snow and ice layers in the snowpack that are hard for the reindeer to penetrate (Riseth et al., 2010). These relate to observations of more frequent winter thaws and rain-on-snow events (Shmakin, 2010) (Figure 4.12) that can have catastrophic consequences for animal populations (see section 4.4.3.2). The observations also relate to heavy snowfall under relatively high temperatures and compression by later snowfall. In Canada, traditional knowledge records evidence of harder snow that impacts the construction of snow shelters (Walser, 2009). In contrast, observations of recent changes in snow stratigraphy for northern Eurasia showed that interannual variability outweighed any possible trends (Golubev et al., 2008).

An ice layer at the bottom of the snowpack ('basal ice') is an important problem for Arctic grazing animals such as reindeer, caribou, musk ox, and small rodents such as lemmings, particularly if the ice layer forms early in the season and restricts access to forage for an extended period (Forchhammer and Boertmann, 1993; Aanes et al., 2000; Solberg et al., 2001; Griffith et al., 2002). Basal ice layers can form at any time during the snow season from thaw events followed by subsequent refreezing. However, the data archive of Bulygina et al. (2010a) suggests that they tend to be more frequently encountered in spring. In Russia, the thickness of the basal ice layer has been routinely measured at 958 long-term stations since 1966 as part of the meteorological observing program (Bulygina et al., 2010a). Dangerous events for reindeer husbandry (DER) are reported by the Russian Meteorological Service when the basal ice layer is thicker than 5 mm over ten consecutive days. A recent analysis of DER events in Russia since 1966 (Bulygina et al., 2010a) found a downward trend of about 5% per decade in the number of sites reporting DER events, mainly in response to a shorter and more intense snow melt season. Further analysis of DER frequency in the early winter period is required to determine if there are changes with potential impacts for grazing animals.

4.3.1.3.2. Observed changes in albedo and snowpack chemistry: black carbon effects

Snow is the most reflective natural surface on Earth, with albedo typically 70% to 80% for the freshly fallen snow. However, albedo gradually decreases with snow ageing and contamination by external materials that accumulate in the snowpack. Nevertheless, the decreased albedo of the older snow cover remains much higher than that of all underlying surfaces except ice. Because the albedo is so high, it can be reduced by even small amounts of dark impurities. Just a few parts-per-billion (ppb) of black carbon (soot) can reduce the albedo of snow by 1–4%, as the black carbon strongly absorbs solar radiation (Warren and Wiscombe, 1980, 1985; Flanner et al., 2007). The absorbed radiation is converted to internal energy that is re-emitted as heat to the surrounding snow or ice and air. Hence, deposition of black carbon and other aerosols, such as deposition from volcanoes, onto snow and ice surfaces can increase melt rates (although the study by Jones et al. (2005) on the climate effects of a 'super volcanic eruption' suggested increasing snow cover in response to the cooling). Snow with black carbon, therefore, melts sooner in the spring and uncovers the darker underlying surface, causing an amplifying feedback on climate (Hansen and Nazarenko, 2004; Grenfell et al., 2009).

Some black carbon emissions result from marine transportation and offshore fossil fuel extraction within the Arctic (Macdonald et al., 2005), activities which are likely to increase as sea ice retreats (see Chapters 9 and 10). However, sources of black carbon are primarily located outside the Arctic. The main source is biomass burning (Hegg et al., 2009); forest fires and agricultural fires as well as the burning of fossil fuels. Nevertheless, black carbon particles can travel long distances and are able to reach the Arctic in substantial quantities. Observations show a large seasonal cycle in atmospheric levels, with maximum concentrations in surface air during late winter and early spring and minimum values in late summer and early autumn (Sharma et al., 2006). Black carbon particles are typically hydrophobic when emitted, but age fairly rapidly to a hydrophilic state by mixing with other particles in the atmosphere. They are then deposited at the surface primarily by wet deposition, although dry deposition also plays a small but important role. Atmospheric residence times are generally about three to eight days (Shindell et al., 2008). A typical mid-latitude snow crystal contains thousands of particles, including absorbing black carbon and mineral dust (Chylek et al., 1987).

Clarke and Noone (1985) measured black carbon in snow throughout the western Arctic. The cleanest snow is in Greenland, with about 2 ppb. Snow in Canada, Alaska, and on sea ice in the Arctic Ocean has 5 to 10 ppb (Hegg et al., 2009), which is less than was reported by Clarke and Noone (1985). This result is consistent with the decline in atmospheric black carbon levels measured continuously at Alert since 1989 (Quinn et al., 2007). Snow in the Russian Arctic sampled in the winters of 2007 and 2008 (Grenfell et al., 2009) showed higher values than elsewhere in the Arctic, typically in the range 20 to 30 ppb. The median background values were 15 to 25 ppb on or near the coasts of the Barents and Kara Seas and near a polynya located within the Laptev Sea. Values were 15 to 20 ppb on the Chukchi Peninsula, but only 5 ppb in a fresh snowfall event at the end of April. Farther south and west, background levels were higher at 20 to 80 ppb in the Sakha Republic and 40 ppb on the Taymir Peninsula (Grenfell et al., 2009).

The change in albedo due to the presence of black carbon can have substantial effects on snow melt and atmospheric temperatures when large amounts of sunlight reach the snow surface – the 20-year global warming potential of black carbon is 2000 carbon dioxide equivalents (Hansen et al., 2007). The effects of black carbon are greatest in large, open areas and in late spring, summer, and early autumn. The albedo reduction also varies with the age of the snow, with older, larger-grained snow showing roughly three times the reduction of new, smaller-grained snow (Warren and Wiscombe, 1985). More broadly, by reducing the reflection of solar radiation to space, deposition of black carbon onto snow and ice leads to warming of the planet as a whole, resulting in increased melt rates of snow and ice. Model studies suggest that black carbon plays a very large role in spring snow melt, with about 20% to 30% greater melting than in simulations that do not include the effects of black carbon deposits (Flanner et al., 2007). Experiments suggest that emissions of black carbon and organic matter from fossil fuel combustion induce 95% as much loss of springtime snow cover over Eurasia as anthropogenic carbon dioxide (Flanner et al., 2008).

4.3.2. Projected changes in snow cover and its characteristics

4.3.2.1. Methodology to predict changes in snow cover and characteristics

The present (2009) ability of the scientific community to provide guidance on how Arctic snow cover will respond to climate change is limited by a number of issues. First, there is no systematic collection of pan-Arctic snow data for monitoring changing snow cover conditions and for developing and evaluating climate models. Second, for computational reasons, the CMIP3 suite of coupled climate models used relatively simple snow schemes that did not include many processes important for high-latitude snowpack evolution (Brun et al., 2008; Holko et al., 2008). In addition, Kattsov et al. (2007) showed that the CMIP3 models tended to overestimate precipitation over major river basins in the Arctic due to inadequate treatment of orography and biases in atmospheric and sea-ice circulation. Räisänen (2007) and Brown and Mote (2009) found that the CMIP3 model mean SWE climatology over the Northern Hemisphere agreed reasonably well with available observations. However, there is also recent evidence that the atmospheric circulation patterns and snow feedbacks in the CMIP3 models are unrealistic over high latitudes. For example, Hardiman et al. (2008) showed that none of the CMIP3 models were able to reproduce the observed strong correlations of Eurasian autumn snow extent to atmospheric wave activity and Northern Annular Mode (NAM) anomalies in the following winter. Also, Fernandes et al. (2009) noted that the CMIP3 models do not properly capture the spatial and temporal characteristics of northern hemisphere snow temperature sensitive regions documented by Groisman et al. (1994a). Inadequate representation of snow-albedo feedbacks may be contributing to this, as previously noted by Qu and Hall (2007).

In light of these limitations, it is unrealistic to expect GCMs to provide high-quality projections of future changes in snow cover over the Arctic region. However, they can provide an indication of large-scale changes in precipitation- and temperature-dependent snow cover variables, such as snow cover start and end dates and maximum winter accumulation, which are important for many applications (see Section 4.3.2.2). Changes in snow properties such as snow density, ice layers, and changes in spatial accumulation patterns require the use of more-sophisticated models and downscaling approaches (discussed in Section 4.3.2.3).

4.3.2.2. Climate model projections of changes in snow cover and snow water equivalent

The response of northern hemisphere snow cover to climate change simulated by the CMIP3 models did not receive much attention in the IPCC Fourth Assessment (Solomon et al., 2007). Subsequently, Räisänen (2007) examined projected 21st century changes in the SWE for 20 of the CMIP3 general circulation models and found that, while the simulated warming shortens the snow season in both autumn and spring in all of Eurasia and North America, SWE at the height of winter generally increased in the coldest areas and decreased elsewhere. Regions with increasing levels of SWE coincided with the position of the -20 °C isotherm in late 20th century November to March mean temperature (Räisänen, 2007) that covers the northernmost portions of both continents (Figure 4.1). Already, there is evidence of increasing snow depth over northern Eurasia (e.g., Kitaev et al., 2005; Bulygina et al., 2009) (Figure 4.6) but less evidence of increased snow accumulation over the Canadian Arctic (Brown and Mote, 2009).

Analysis of the model consensus pattern for change in maximum SWE (Brown and Mote, 2009) (Figure 4.13) showed that the response consisted of three broad zones: with significant decreases at lower latitudes; a broad zone over mid-to-high latitudes, where changes were not statistically significant; and the northern zone of increasing levels identified by Räisänen (2007). The temporal evolution of the climate change response of SCD (Figure 4.14) is different in that decreases dominate (no regions were identified where climate models show significant increases in SCD), with the earliest and largest decreases in SCD occurring in coastal regions of the continents. Taken together, these figures suggest different regional snow cover responses over the Arctic, with the largest and most rapid decreases both in SWE and SCD over Alaska, northern Scandinavia, and the Pacific coast region of Russia.

Projected changes in maximum monthly SWE (SWEmax) and annual SCD for the 2049–2060 period versus 1970–1999 are provided in Figure 4.15 from a composite of six of the highest resolution GCMs that met the Arctic performance criteria applied in Chapter 3. The choice of six models was based on the findings of Chapter 3 that composites formed from the best-performing five to seven models agreed more closely with observations than composites formed from fewer or more models. The six models used are CCSM3, CNRM, ECHAM5, GFDL, HADGem1, and MIROC32, and results are computed for the IPCC A2 emissions scenario. The projections suggest slight increases in SWEmax (0–15%) over much of the Arctic, with the largest increases (15–30%) over the Siberian sector. Annual SCD is projected to decrease by about 10–20% over much of the Arctic, with the smallest decreases over Siberia (< 10%) and the largest decreases over Alaska and northern Scandinavia (30–40%). The climate models project similar decreases in snow cover at the start and end of the snow season (not shown). The climate model standard deviations (right-hand panels in Figure 4.15) show relatively high values of model consistency (standard deviations <10%) over the Arctic region both for SWEmax and annual SCD with areas of lower model consistency over Alaska and western Europe. The process of interpolating climate model output with different resolutions to a standard grid contributes to higher model standard deviations in coastal mountain areas.

Some aspects of these climate model snow-cover change projections may not be realistic. Brown and Mote (2009) found no evidence in the climate models of the accelerating reduction in high-latitude spring snow cover documented by Déry and Brown (2007) consistent with polar amplification of warming and an enhanced albedo feedback in spring. The muted spring response in the models could be due to a number of reasons, including lack of black carbon darkening (Flanner et al., 2008), inadequate snow-albedo treatments (Qu and Hall, 2007), and wet-cold biases in models over high latitudes (Randall et al., 2007). The slower decrease in SCD over eastern Eurasia shown by the climate models is also inconsistent with observed trends and may be linked to difficulties in simulating the climate interactions of the Tibetan Plateau (Cui et al., 2007).

Snow cover changes are also likely to be much more complex in mountainous terrain than represented in the coarse-resolution GCM results. A sensitivity analysis of snow cover changes to increasing temperature and precipitation suggested a potentially complex elevation response of snow cover in mountain regions due to non-linear interactions between the duration of the snow season and snow accumulation rates (Brown and Mote, 2009). This non-linear response can be expected to contribute to regional-scale variability in the elevation response of snow cover to climate change, which will be modified by local factors such as lapse rate (rate of change with increasing altitude), aspect, topography, and vegetation.

4.3.2.3. Snow cover information for adaptation and decision-making – downscaling approaches

For many applications, the snow-cover change information from GCMs is too coarse for use by decision-makers. There is a wide range of options for downscaling information to the required level, with an extensive body of literature and guidelines (e.g., Wilby et al., 2004; Christensen et al., 2007; Benestad et al., 2008). The two main approaches are dynamical and statistical downscaling. Dynamical downscaling uses higher resolution regional climate models (RCMs) driven with lower resolution GCM data as the boundary forcing. This approach has been used in Norway (Vikhamar-Schuler et al., 2006), the western United States (Rauscher et al., 2008), and Switzerland (Bavay et al., 2009) to investigate the local and regional response of snow cover to different climate change scenarios. These high-resolution model-based approaches are computationally expensive to run but are attractive in that they provide a physically based approach to downscaling (although care must be taken to account for any biases in RCM fields that can be passed from the driving model or generated by the RCM, such as in the land surface scheme).

A number of high-latitude environment distributed process models have been developed in recent years (e.g., Liston et al., 2007; Pomeroy et al., 2007; Schramm et al., 2007; Andreadis et al., 2009; Shmakin et al., 2009) that can be driven with atmospheric re-analyses or RCMs to provide assessments of variability and change in snow cover and hydrology at basin to pan-Arctic scales (e.g., Su et al., 2005). Previous

evaluations of snow models in the Arctic (Slater et al., 2001; Bowling et al., 2003) have shown that most models are able to provide reasonable snow cover simulations given good quality input data. The challenge is to provide realistic driving variables, particularly precipitation, to these process models. The potential for using snow information directly from existing re-analyses such as ERA-40, NCEP, and JRA-25 is limited by the relatively coarse resolution and because the snow fields must be carefully evaluated on a regional basis due to spatially and temporally varying biases and errors in the assimilation of snow observations (Khan et al., 2008). The Arctic System Reanalysis project (an IPY Project) is intended to address a number of these issues (http://polarmet.mps.ohio-state.edu/PolarMet/ASR.html).

There are few pan-Arctic RCMs under development, and their initial evaluation (Rinke et al., 2006) showed large scatter among models and reduced confidence in air temperatures over land, surface radiation fluxes, and cloud cover. More recently, Ma et al. (2008) carried out a five-month climate simulation with the modified mesoscale model, Polar MM5, coupled to the NCAR Land Surface Model (LSM) and found that the coupled model improved forecast skill for surface variables at some sites. Ongoing work developing the Polar Weather Research and Forecasting (Polar WRF) model (e.g., Hines and Bromwich, 2008) is providing guidance on the most appropriate physical parameterizations for use in polar regions. Three climate model intercomparison projects (MIP) are in progress to improve Arctic climate models: the Arctic Regional Climate MIP (ARCMIP), the Arctic Ocean MIP (AOMIP), and the Coupled Arctic Regional Climate MIP (CARCMIP) for coupled atmosphere-ice-ocean-land models.

Statistical downscaling techniques are based on the development of statistical relationships between low-resolution GCM fields and the high-resolution observed records (usually surface variables such as precipitation or temperature) through a variety of methods, including regression analysis, weather typing, and analogues (see Kattsov et al., 2005; Benestad et al., 2008). While there are few examples of statistical downscaling applied directly to snow cover in the Arctic, there are many examples of downscaling temperature and precipitation information (e.g., Kattsov et al., 2005), which are two of the basic ingredients for simulating snow cover.

The development of scenarios for changes in snow cover properties relevant to ecological studies (e.g., snowpack structure, ice layers, rain-on-snow events) will require the use of detailed physical models that take account of snow layering. There are several such models available (e.g., CROCUS, Brun et al., 1989; SNTHERM, Jordan, 1991; and SNOWPACK, Bartelt and Lehning, 2002), and these can be readily run in 1-D column mode at individual sites with locally downscaled driving fields from climate models. However, running these models in a spatially distributed mode like Bavay et al. (2009) is computationally expensive, and ongoing work is needed to find ways to efficiently couple detailed snow process models into distributed Arctic process models that take account of heterogeneous terrain and vegetation (Brun et al., 2008). The recent simplified five-layer snowpack model developed by Andreadis and Lettenmaier (submitted) may be a suitable compromise, as it was able to provide realistic simulations of Arctic snowpack structure along a transect in northern Alaska. An example of snow depth projections using the HIRHAM RCM is shown in Figure 4.16

4.4. Evaluation of the impacts of a changing snow cover and its characteristics

- Changing Arctic snow climate has had widespread impacts and future changes will probably intensify these. Impacts can be beneficial or deleterious depending on the snow cover sensitivities involved.
- Diminished spring floods and a longer snow-free season increase the potential for evaporation and threats to wetlands, but this may be offset by new sources of water from thawing permafrost and associated thermokarst. The critically important balance between the two processes is unknown and areas of drying tundra as well as of increased water logging are being reported across the Arctic.
- Snow-vegetation interactions are expected to play an important role in the future evolution of northern hydrology and thermal regimes. The enhanced growth of Arctic shrubs in response to warming increases the potential for local increases of snow cover accumulation independent of increases in precipitation. Increased

shrubiness will also change snowpack properties and melt dynamics that in turn will have implications for the soil thermal regime and hydrology.

- Reduced snow cover duration shortens the winter snow transportation season, which is important for many Arctic communities and commercial enterprises.
- Snow-related hazards could increase in areas with projected increases in annual maximum snow accumulation.
- Increased winter precipitation and a more even seasonal distribution of runoff may prove beneficial to the hydropower industry.
- A shorter snow cover duration is likely to increase plant productivity and carbon capture where soil moisture is adequate.
- Reductions in the amount of water stored in the winter snowpack are already increasing the moisture stress of northern coniferous forests in summer.
- Winter thaws and rain-on-snow events that are projected to increase in frequency are already damaging vegetation and Arctic grazing mammal populations and will lead to additional challenges for commercial reindeer herding.

4.4.1. Changes in snow cover and impacts on climate processes

Snow cover plays a major role in climate, and the hydrological and ecological systems through its influence on the surface energy balance (e.g., reflectivity), water balance (e.g., water storage and release), thermal regimes (e.g., insulation), and vegetation and trace gas fluxes (see Figure 4.17).

The main snow-climate processes are briefly presented here in Section 4.4.1, while the impacts of snow on the hydrological cycle, vegetation, trace gases, and the accumulation of black carbon that affect the climate system are presented in Sections 4.4.2, 4.4.3.1, 4.4.3.3 and 4.3.13.2.

4.4.1.1. Changes in surface boundary layer processes, such as atmospheric inversions

Temperature inversions close to the ground surface are characteristic features of Arctic climate, particularly in winter when the ground is snow covered. Stable conditions can persist for weeks, decoupling the surface from atmospheric conditions a few kilometres aloft. They thus affect the surface energy exchange. Low-level inversions result from two basic conditions:

1. A radiative imbalance, in which surface energy emission exceeds that received directly from solar radiation and the atmosphere (a condition that is common in the Arctic snow season, when net radiation is generally negative, and during the rest of the year, when the sun is lowest in the sky).

2. Warm air advection over a cooler surface (common in snow-covered regions and over sea ice), a condition which may occur at any time of year. During late spring and early summer, this may involve the formation of shallow surface inversions with only slightly cooler temperatures near the surface as melting snow and ice act as a heat sink. However, in winter, warm air aloft can create extremely stable conditions with temperature gradients in the lower troposphere of more than +6 °C/100 m (Overland and Guest, 1991; Bradley et al., 1992a).

Studies of inversion characteristics over time show a widespread decline in surface-based inversion depth from the 1950s to the 1980s across Arctic Canada and Alaska (Bradley et al., 1992b; Hartmann and Wendler, 2005; Bourne, 2008). However, over the past 20 years inversion depths have shown no further downward trends at most sites. It is possible that some of the declines seen in the early period were artifacts of improved instrumentation (faster response times and thus improved detection of atmospheric structure; see Walden et al., 1996), but the widespread and steady nature of the observed declines in Siberia, Alaska, and Canada make such an explanation unlikely to be applicable in all areas. Circulation changes (such as the mid-1970s shift in the Pacific Decadal Oscillation) appear to have had a strong influence on inversion structure in Alaska (Bourne, 2008). Circulation changes across the Arctic have

contributed significantly to warming (Graversen et al., 2008), and it is likely that such changes have affected the near-surface inversion structure.

General circulation models perform poorly at simulating conditions in the near-surface boundary layer and there have been few attempts to use regional models to dynamically downscale GCMs over the Arctic. However, initial results indicate that such simulations overestimate surface temperature and daily inversion variability, and underestimate inversion depths and strength (Bourne, 2008). Simulations of future winter climate in Alaska (for emissions scenario A1B) downscaled to the regional level for Alaska (MM5 driven by CCSM3) suggest that inversions will become shallower and less strong (Bourne, 2008), but given the relatively poor performance of these models in simulating modern conditions, these results can only be considered as suggestive rather than definitive. Much additional research using different regional models is needed before a clear picture of how anthropogenic global warming may be expressed in terms of the surface-based inversion structure in the Arctic.

4.4.1.2. Changes in the atmospheric moisture budget in relation to snow cover

The atmospheric moisture budget is changing and will continue to evolve throughout the 21st century in response to a warming climate, changes in sea ice and snow cover, and altered atmospheric circulation (Rawlins et al., 2010). Increased winter snowfall over land areas may act to keep soil moisture levels higher in spring, working with increased precipitation to promote more evaporation. On the other hand, under a warming climate earlier loss of the snow cover over Arctic and sub-Arctic lands, by exposing the dark underlying surface, will promote surface heating, and hence, an earlier transition to the summer-type convective precipitation regime over Arctic and sub-Arctic lands (Groisman et al., 1994b).

Simulations from most coupled GCMs point to intensification of the northern high-latitude freshwater cycle through the 21st century (e.g., Holland et al., 2006; Kattsov et al., 2007). A warmer atmosphere carries more water vapor and with efficient precipitation-generating mechanisms (convergence and uplift), precipitation increases. While evaporation also increases, the precipitation change dominates such that there is an increase in net precipitation and hence river discharge. Although the balance of observational evidence points to intensification over the past several decades, intrinsic variability and lack of consistency in observed trends limit confidence in the robustness of the changes (Rawlins et al., 2010).

A warmer Arctic atmosphere also provides more energy for sublimation. A review of the blowing snow sublimation process was provided by King et al. (2008) with an estimate of ~20% of annual snowfall for Arctic blowing snow sublimation losses. Providing estimates of how this may change in the future is complicated by difficulties in scaling-up the local-scale processes that drive the blowing snow sublimation process (King et al., 2008).

Although the future is uncertain, there are likely to be changes in patterns of atmospheric circulation affecting pathways of moisture transport and in turn influencing precipitation and snow cover. Some simulations suggest that in the future there will be a more frequent positive phase of the Northern Annular Mode (NAM) and its Atlantic-side component, the NAO (Rogers and van Loon, 1979). How altered precipitation linked to future behavior of the NAM and NAO and other patterns of atmospheric variability translates into changes in snow cover will depend on attendant temperature rise. This translation is important given the potential for snow cover forcing on circulation. For example, there is evidence from models (e.g., Walland and Simmonds, 1997) and observations (Clark and Serreze, 2000) that variability in Eurasian snow cover extent modulates atmospheric circulation patterns over the North Pacific Ocean. In addition, the recent modeling study of Deser et al. (2010) suggests that a seasonal cycle of sea-ice extent projected for the late 21st century will promote increases in winter snow depth over Siberia and northern Canada because of increased winter precipitation.

4.4.2. Changes in the hydrological cycle at the local and regional scale

4.4.2.1. Regional and seasonal variability in sources of moisture for precipitation

Within the Arctic drainage basin, maximum monthly precipitation occurs in the summer to autumn period, but the maximum discharge is observed in June, when it primarily originates from snow accumulated over the long (five- to eight-month) cold period (Figure 4.18). Long-term averages of precipitable water calculated from the atmospheric moisture budgets of the large Arctic-draining Eurasian watersheds (Ob, Yenisey, Lena, Kolyma-Indigirka) have symmetric annual cycles, with July peaks and winter minima (Serreze and Etringer, 2003). This reflects the annual cycle in atmospheric temperature and the ability of the atmosphere to carry water vapor. Precipitation also has a symmetric annual cycle with a summer peak. While effective precipitation-generating mechanisms exist in all seasons, these watersheds are far removed from oceanic moisture sources ('continentality'). As such, precipitation tends to follow the seasonality in the available column water vapor. During winter, the primary precipitation mechanism is a modest convergence of water vapor; this precipitation is stored in the winter snowpack and released in spring and summer as river discharge.

Processes over the major Eurasian catchments contrast sharply with the Atlantic sector of the Arctic, which has a general cold season precipitation maximum. Here, the annual cycle of precipitation is not moisture limited but rather reflects the stronger precipitation-generating mechanisms in the cold season (i.e., vapor flux convergence and uplift associated with eddy activity along the North Atlantic storm track). In this region, the annual cycles of precipitation and column water are in opposition (Serreze and Etringer, 2003).

In the Arctic Ocean region, precipitable water has a July peak and winter minimum (AARI, 1985). Similar to the terrestrial watersheds, the vapor flux convergence tends to have a general cold season minimum and summer to early autumn peak (e.g., Walsh et al., 1994; Yang, 1999). Here, however, evaporation is always low, limited in winter by low temperatures and in summer by the presence of a melting snow and sea-ice surface. For all seasons, precipitation is primarily related to the horizontal vapor flux convergence.

4.4.2.2. The role of snow cover in the seasonal hydrological cycle

Snow cover has a direct impact on the hydrological cycle in its redistribution of water between cold and warm seasons, with limited water availability during the cold season and an abundance of water during snow melt. The snow-generated runoff in the Arctic drainage basin is up to 75% of total annual flow in some northern regions of Siberia and North America (Woo, 1980).

Whereas snow cover has a direct impact on the hydrological cycle in its redistribution of water between cold and warm seasons, the indirect snow cover impact has its effects on the surface energy budget. That is, on the latent heat flux to the atmosphere (suppressed by low surface temperatures associated with snow cover and the absence of transpiration) and to the soil (by intercepting precipitation and meltwater and sending a significant part downstream during the snow melt period instead of contributing to the baseflow; Dingman, 2002). Therefore, future variations in snow depth may dramatically change the hydrological regime of the Arctic (Adam et al., 2009).

Changes in precipitation and temperature are expected to affect all aspects of the cryospheric hydrological cycle, including snow accumulation and melt, evaporation, and runoff as well as short- and long-term storage changes. Net snow accumulation at the end of winter will be modified by winter warming and precipitation enhancement in the circumpolar areas, but their effects would vary between regions (see Section 4.3.2). In other Arctic and sub-Arctic areas, projected increases in winter precipitation may be accompanied by increased occurrences of winter thaw. Mid-winter melt events and rainfall instead of snowfall can lead to a reduction in net snow accumulation, and such tendencies are suggested by the records of some stations in northwestern Canada (Burn et al., 2004). Changes in the physical and biological environment further influence winter snow accumulation. The polar sea ice is retreating at a drastic rate, and this shortens the ice-covered season while creating or enlarging polynyas (areas of open

water surrounded by sea ice) in the Arctic Ocean. The Svalbard islands offer a present-day analogue for many Arctic islands under scenario climates. Humlum et al. (2003) noted that when airflow over open waters advects warm air to the islands in the winter, it is often accompanied by heavy snowfalls and periods of snow melt.

Shortening of the snow cover period leads to a corresponding extension of the evaporation season. Earlier snow melt under a future climate will be accompanied by an earlier breakup of lake and river ice (see Chapter 6)

Changes in streamflow will be expressed mostly through high and low flows. At present, spring floods usually represent the highest flow of the year, generated by rapid melting of a snowpack built up in the long winter. A shortened snow accumulation season interrupted by mid-winter melt events will diminish the water supply and moderate its rate of delivery. An increase in groundwater discharge in winter, however, can produce more icing (also known as 'aufeis' or 'naled') in stream channels to block the spring flow and intensify the snowmelt flood locally. Streamflow data from northern Canada show that there are detectable trends of decreasing freshet (a sudden overflow of a stream from a heavy rain or thaw) peaks and earlier arrival of spring floods (Woo and Thorne, 2003; Burn et al., 2004; Aziz and Burn, 2006).

After the spring flood, streamflow declines to its summer low. With an extended snow-free period and higher evaporation rates, groundwater drawdown will be accompanied by reduced baseflow. Small basins with polar desert conditions may find a total cessation of their summer flow except during heavy rains. Recent studies suggest that there may be a link between significant increases in winter baseflow of many northward-flowing rivers and permafrost thawing (St Jacques and Sauchyn, 2009). Plausible causes include a thicker active layer or forest fire or precipitation increase in the cold months (McClelland et al., 2004; Rawlins et al., 2009; Shiklomanov and Lammers, 2009). However, there is incompatibility between positive trends in the runoff of the Yenisey and Lena Rivers and the lack of positive trends in precipitation data (Figure 4.18), perhaps owing to data quality issues (Berezovskaya et al., 2004). Human activities such as reservoir operation also have overwhelming effects on streamflow regimes (Ye et al., 2003; McClelland et al., 2004; Yang et al., 2004; Woo et al., 2008).

Although snow is currently the principal promoter of river flow, in future the number and magnitude of rainfall events may increase, while the contribution of snow may decline, notably in the sub-Arctic. The seasonal flow pattern may then be replaced by a pluvial or a nivo-pluvial regime. Under climate change scenarios of warmer winters and more rain events, simulations of streamflow in mountainous basins of northern Canada suggest that winter flow will increase and spring freshet dates will advance (Figure 4.19a), but peak flow will decline, as will summer flow due to intensified evaporation (Kerkhoven and Gan, 2005; Woo et al., 2008). Thus, the total annual flow is likely to remain the same (see superimposed present and future probabilities in Figure 4.19b). An earlier melt season will spread the basin snow melt over a longer period (Figure 4.19c, d, e) (Woo et al., 2008).

Reduced snow accumulation, earlier snow melt, and increased evaporation (and sublimation) have consequences for seasonal and longer-term storage of water and ice. Many late-lying and semi-permanent snowbanks will shrink or disappear from the Arctic landscape. Snow storage on glaciers will change with negative consequences for their mass balance (see Chapter 7). The loss of snowbeds will lead to the demise of many patchy wetlands in the Arctic that depend on the late-lying snow to sustain wetland saturation (Woo and Young, 2003). For all northern wetlands, snowmelt flood is of great importance, as it replenishes wetland storage every spring through infilling of surface depressions and saturation of the peat and mineral soils (Woo and Guan, 2006). With diminished spring floods and a longer snow-free season for evaporation, the hydrological and ecological health of northern wetlands could deteriorate, but thawing permafrost and associated thermokarst could provide a new source of water. The critically important balance between the two processes is unknown, and areas of drying tundra as well as of increased waterlogging are both being reported (see Chapter 5).

4.4.3. Ecological processes and the role of snow cover

4.4.3.1. Interactions between snow and vegetation

The interaction between snow and vegetation is complex, and can involve interception, sublimation, snow trapping (Figure 4.20), energy exchanges, and hydrology. On the one hand, vegetation has strong and complex influences on the accumulation and ablation of snow (Pomeroy et al., 2006), which are expected to alter as vegetation distribution or structure change under a warming climate. On the other hand, snow provides protection for vegetation from extreme low temperatures, large temperature fluctuations, ice crystal blast, and desiccation.

4.4.3.1.1. Effects of vegetation on snow – vertical profile

Snow falling onto shrubs and trees is partitioned into interception by the canopy and throughfall to the ground (Hedstrom and Pomeroy, 1998). As the intercepted snow load increases, the interception efficiency increases due to snow bridging between canopy elements but decreases due to bending of branches under the load. Canopy capacities can be much greater for snow than for liquid water and intercepted snow can remain in a forest canopy for some time. Snow can be removed from canopies by direct unloading, drip of meltwater, and sublimation (Storck et al., 2002; Molotch et al., 2007).

According to Sámi reindeer herders, large trees in old forests trap part of the snowfall, thereby reducing the depth of the snow on the ground. In younger stands, with only small seedlings, most of the snow accumulates on the ground, resulting in a deeper snow cover (Kumpula et al., 2007). When all the snow that has accumulated in the canopy of a dense even-aged stand melts and falls, the snow cover is compacted over the whole stand. In such cases, reindeer herders can only use stands with low stem density or stands in which younger trees with smaller crowns have not trapped as much snow (Roturier and Roue, 2009).

4.4.3.1.2. Effects of vegetation on snow – horizontal aspects

Vegetation reduces the horizontal redistribution of snow and can therefore have greater snow depths than open areas, particularly downwind of vegetation patch edges where snow relocated from open areas can be trapped. Snow trapping depends on the height, density, and distribution of vegetation, all of which are projected to change under a warming climate. Snow in shrub patches contains a greater percentage of low-density depth hoar, increasing the insulation of underlying soil (Sturm et al., 2001a).

4.4.3.1.3. Effects of vegetation on the energy balance of snow

Vegetation can increase or decrease snow-melt rates relative to open areas depending on vegetation characteristics (Davis et al., 1997; Lee and Mahrt, 2004; Pomeroy et al., 2006). However, many processes are involved, and the net outcome in the context of vegetation changes is difficult to model.

Vegetation shades snow on the ground from solar radiation to an extent dependent on the density of canopy elements exposed above the snow and the fractions of the incoming radiation that are direct or diffuse. In areas of discontinuous vegetation cover, even snow in open areas can be shaded by cast shadows, particularly in the Arctic where the sun is low in the sky. Although sophisticated models of radiative transfer through vegetation canopies exist (e.g., Ni et al., 1997), practical applications almost invariably use simpler models that only consider vertical fluxes and thus cannot represent shading or canopy gaps (Bewley et al., 2007); the measurement of accurate spatial statistics is also challenging (Link et al., 2004).

Merely because the thermal emissivity of canopy materials is higher than that of air, the presence of a vegetation canopy can greatly increase the net thermal radiation absorbed by snow (Sicart et al., 2004).

Moreover, because vegetation has a lower albedo than snow and can be warmed to temperatures above 0 °C, sun-lit vegetation can have temperatures well in excess of the air and snow temperatures, further increasing the thermal radiation emitted to the snow (Pomeroy et al., 2009). These influences will be least under dense vegetation that limits the penetration of solar radiation and greatest close to the sun-lit edges of vegetation patches (Essery et al., 2008). Because snow generally has a high albedo and only absorbs a small fraction of the incoming solar radiation, the increase in thermal radiation due to the presence of vegetation can exceed the corresponding reduction in solar radiation in the radiative energy balance of a snowpack (Sicart et al., 2004). Sparse vegetation debris lying on a snow surface reduces its albedo (Hardy et al., 1998), thereby increasing melt rates while buried branches and litter absorb solar radiation that can penetrate to some depth in snow, possibly leading to subsurface melting. Turbulent transfers of heat to a snow surface are increased by the exposure of sparse vegetation but decreased by dense vegetation cover (Reba et al., 2009).

Ground vegetation also plays a key role in thawing and freezing events. According to Sámi reindeer herders, in forests with a thick layer of vegetation, the water flows down through the generally thick moss layer, leaving the lichen accessible for the reindeer to graze. In contrast, in thinner layers of lichen-dominated vegetation, the water soaks into the lichen mat and freezes, forming a crust of ice and limiting its availability as reindeer fodder (Roturier and Roue, 2009) (see Section 4.4.3.2 for impacts of these freezing and thawing events on animals).

4.4.3.1.4. Effects of snow on vegetation

There is a close relationship between snow distribution and vegetation type (Evans et al., 1989), and the heterogeneity of the snow regime helps drive the (large-scale) diversity of the Arctic flora. For example, the high wind speeds in the Arctic ensure considerable redistribution of snow, and some areas may be exposed and snow free for much of the winter, while others, such as snowbeds that often have relatively high diversity and harbor rare plant species, accumulate snow. Exposed areas typically have cushion plants, evergreen dwarf shrubs and tussock graminoids, whose forms provide protection from wind abrasion and summer drought stress (Walker et al., 2001). Plants covered by snow in winter often have the capacity to burst bud quickly after snow melt from buds pre-formed in the previous growing season and using carbohydrate reserves from the previous season's photosynthesis. Such rapid response maximizes their use of the short Arctic growing season (Chambers, 1991; Walker et al., 2001) determined by the period and timing of SCD. The ongoing and projected changes in snow conditions presented in earlier sections will profoundly affect the distribution of vegetation and biodiversity.

Arctic plants may be physiologically active under snow before completion of melt, especially in the later stages of melt when soils may be warmer and enough light may penetrate the snow to allow photosynthesis. Lichens and bryophytes are most capable of this, as they are physiologically active at low temperatures, have low-light compensation points, are able to acquire moisture and nutrients directly from meltwater, and have an opportunistic growth strategy that allows quick reactivation when suitable growth conditions occur (Kappen et al., 1995; Walker et al., 2001; Kappen and Valladares, 2007). However, such activity is detrimental to lichens, because their freeze tolerance has a high carbohydrate cost, particularly in warm, low-light subnivean cavities (Kappen et al., 1995). An increase in SCD or warm subnivean conditions may result in declines in these species (Benedict, 1990).

In contrast, vascular plants (e.g., herbs, shrubs, grasses, sedges) in the Arctic show less ability for subnivean activity, although they are capable of rapidly increasing rates of photosynthesis following snow melt (Tieszen, 1974; Defoliart et al., 1988) and some vascular plants are capable of net photosynthesis below snow during melt (Semikhatova et al., 1992; Starr and Oberbauer, 2003). Such favorable conditions may last for two to four weeks and thus represent a significant extension of the growing season. Indeed, cold season photosynthesis (from vascular and non-vascular plants) was recently found to constitute a considerable amount (19%) of the gross annual CO_2 uptake in sub-Arctic heathland (Larsen et al., 2007a).

Snow can provide nutrients to some plants. Arctic vascular plants captured less than 0.1% of their annual nitrogen requirement from melting snow in the tundra (Bilbrough et al., 2000), while bryophytes and lichens gained ten-fold more. Similarly, in the High Arctic, bryophytes and lichens (as well as soil microbes) showed much better ability to acquire snowpack nitrogen than vascular plants (Tye et al., 2005).

Snow manipulation experiments suggest that responses of vegetation to changing snow-melt timing may not be simple. Some snow-depth manipulations suggest that where areas have rapid snow melt, even considerable increases in snow depth may have only modest impacts on growing season length (Walker et al., 1999). Nevertheless, delayed snow melt can delay plant phenology and ultimately productivity (Walker et al., 2001). However, delayed vegetative growth and flowering may also lead to delayed senescence and ultimately no increase in productivity (Walker et al., 1999). Similarly, species showing earlier physiological activity induced by snow removal and warming manipulations can show earlier senescence, again with no increase in productivity (Starr et al., 2008). Therefore, where Arctic plants have such determinate growth, an earlier growing season may not mean a more productive growing season, but in the longer term, when species with more southern distributions invade, the system is likely to become more productive (Sitch et al., 2008; Wolf et al., 2008).

4.4.3.1.5. Impacts of winter thaw on vegetation

Projected increases in Arctic winter precipitation may be accompanied by increased occurrences of winter thaw. Melting of snow cover in the Arctic begins at below-freezing air temperatures (it starts with temperatures as low as -5 °C) owing to the influence of solar radiation (Radionov et al., 1996). The first snow melt can occur in winter, and for several months thereafter the snowpack remains on the ground. However, even the first such melt initiates a process of snow metamorphosis on its surface that changes snow albedo by generating a snow crust on the surface and ice layers within the snowpack or at the ground surface. Such crusts and ice layers remain until complete snow melt and have important, often devastating, consequences for some animal populations including domesticated reindeer (see Section 4.4.3.2).

Increases in mid-winter melt events and rainfall instead of snowfall are suggested by the records of some stations in northwestern Canada (Burn et al., 2004) and in the western part of northern Eurasia (Groisman et al., 2003). For example, in Fennoscandia, in the second half of the 20th century, the number of days with winter thaw increased by six over 50 years, or by 35% (Groisman and Soja, 2009; Groisman et al., 2010), while the number of days with rain-on-snow events increased by at least three (see also Figure 4.12; Shmakin, 2010). These events result in sudden loss of snow protection to vegetation. Temperatures rise rapidly to well above freezing, causing snow melt at landscape scales (Robinson et al., 1998; Phoenix and Lee, 2004; Bokhorst et al., 2008, 2009), warming plants and soils and then (following a few days of warming) exposing the ecosystem to rapidly returning extreme cold. Simulation of such events in the sub-Arctic together with observations of a natural event in winter 2007/08 has shown that shrub species may suffer with increased mortality of buds and shoots, delayed bud burst in spring, and reduced flowering and berry production (Bokhorst et al., 2008, 2009). The large scale of the natural event (reduced Normalized Difference Vegetation Index (NDVI) related to productivity over more than 1400 km^2) suggests that extreme warming events may reduce productivity of Arctic vegetation and counterbalance (Bokhorst et al., 2009; Figure 4.21) the long-term trend of shrub expansion into the tundra (Sturm et al., 2001b; Tape et al., 2006).

In the northernmost coastal rainforest region of British Columbia and southeast Alaska, winter thaw events are an important factor in the decline of one of the most valuable timber species in North America: Alaska yellow-cedar (*Chamaecyparis nootkatensis*). This tree has experienced several waves of death over an area of more than 200 000 ha caused by freezing injury to the roots of the trees after the premature loss of winter hardiness (physiological resistance to freezing injury). In recent warm winters, the Alaska yellow-cedar has encountered thawing events in late winter (February) warm enough to cause it to lose winter dormancy (Hennon et al., 2006). When temperatures return to the below-freezing levels normal for

that time of year, freezing penetrates into the snowless ground, killing the shallow roots (Beier et al., 2008). Consequently, Alaska yellow-cedar is being eliminated quickly from the southern and lower elevation portions of its current distribution, while it gradually migrates, becomes established, and matures in the northern and higher elevations beyond its current distribution. The result is the net loss of commercial benefits from the species for a number of centuries.

4.4.3.2. Interactions between snow and animals

Snow is probably one of the most important climatic drivers affecting Arctic biology and its environment (Callaghan, 2005; Meltofte et al., 2008; Post et al., 2009). The interaction between snow and Arctic animal species is determined by the varying abilities of Arctic species to migrate (Figure 4.22; Forchhammer et al., 2008; Robinson et al., 2009). Migratory caribou, for example, prefer regions with higher snowfall and lichen availability during their autumn and winter migrations (Sharma et al., 2009). However, ice crust formation can obstruct migrations. In the Yamal Peninsula, ice crusts may become as common in the northern part as they currently are in the southern part. Such a development would further constrain the possibility for reindeer to migrate on the peninsula (Bartsch et al., 2010).

Whereas species resident in the Arctic (lemming, musk ox, stoat) and locally migrating (fox, caribou/reindeer, ptarmigan) will experience varying snow conditions throughout the year, long-distance migrating species (waders, geese, gulls, skuas, terns) are influenced only by the conditions when they arrive at their Arctic breeding grounds in spring. A 10% decrease in snow cover, advances egg laying by the pink-footed goose (*Anser brachyrhynchus*) by five to six days, corresponding to a 20% increase in probability of nesting success (Madsen et al., 2007). Such considerable responses in population dynamics to changes in snow cover make species highly sensitive to alterations in climate. For the pink-footed goose population breeding on Svalbard, a 1–2 °C increase in temperature that would probably affect snow cover could potentially double the breeding range there (Jensen et al., 2008), provided that conditions at wintering grounds outside the Arctic remain favorable (Wisz et al., 2008). Thus, climate effects across the entire range of a species become pivotal in estimating effects taking place in the Arctic (Figure 4.22; Forchhammer et al., 2008).

In addition to the apparent direct negative effect of increased snow cover on the populations of resident Arctic species such as musk ox (*Ovibos moschatus*) (Figure 4.23a), often mediated through increased over-winter mortality, changes in snow cover may also interact with animals in an indirect and delayed manner through interactions across trophic levels in Arctic ecosystems. Changes in snow cover determine the length of the Arctic growing season (Schmidt et al., 2006; Meltofte et al., 2008) and, hence, the resource base for Arctic herbivores (Forchhammer et al., 2005). For example, over the period 1970 to 2006, reindeer calf production in Finland increased by almost one calf per 100 females for each day of earlier snow melt (Turunen et al., 2009). Similarly, the decadal increase in spring temperature and advance of spring snow melt and consequently the longer growing seasons of 1996 to 2005 observed in northeast Greenland have had a positive influence on the musk ox population there, but the effects of a long growing season were delayed by one year (Figure 4.23b). The relative importance of direct and delayed effects on Arctic herbivores such as the musk ox may vary across the Arctic depending on local climate conditions (Forchhammer et al., 2008). At Ellesmere Island, years with extensive snow cover delayed the vegetation growth, resulting in a halving of the herbivore nutrition-replenishment period, a drastic drop in musk ox and Arctic hare (*Lepus arcticus*) numbers, and a cascading reduction in the population of their predator, the wolf (*Canis lupus*) (Mech, 2004).

Changes in the micro-topical snow conditions may also influence the dynamics of northern rodents (Hörnfeldt et al., 2005; Kausrud et al., 2008). A recent comprehensive study by Kausrud et al. (2008) demonstrated that increased humidity and temperature are related to increased hardiness of the snowpack, which decreases the available subnivean space, causing a decline in rodent population growth rates. A collapse in rodent population size has cascading effects on predators such as Tengmalm's owl (*Aegolius funereus*) (Hörnfeldt et al., 2005). As some predators have other prey in addition to the rodents, such as

ground-nesting birds, the effect of the micro-topical changes in snow conditions that affect the rodent population dynamics could be related to changes in the populations of ground-nesting bird species (Kausrud et al., 2008). Indeed, since any Arctic animal is part of a community and, hence, interacts with other species within as well as across trophic levels, the interaction between snow and the performance of one species may potentially extend to other species (Figure 4.24; Forchhammer et al., 2008; Callaghan and Johansson, 2009; Post et al., 2009).

In addition to the effects of snow depth, SCD, and timing of the snow season on animals, the texture of snow and extreme events in winter are also critical. In extreme cases, crusts, ice layers, and rain-on-snow events have been implicated in population crashes of reindeer on Svalbard (Aanes et al., 2000; Kohler and Aanes, 2004) and the near extirpation of Peary's caribou (*Rangifer tarandus pearyi*) in High Arctic Canada (Barry et al., 2007). On Banks Island in October 2003, a rain-on-snow event is estimated to have contributed to the deaths of 20 000 musk ox by creating a thick layer of ground ice that made foraging very difficult (Rennert et al., 2009). On the Queen Elizabeth Islands, Canadian High Arctic, major population crashes in caribou have been mainly due to exceptionally unfavorable snow or ice conditions (Miller and Barry, 2009).

The Sámi and Nenets have experienced the increasing frequency of extreme weather events. Nenets reported one rain-on-snow event in which impenetrable ice layers formed during a major two-stage thawing and refreezing event over a 48-hour period across an estimated 60 × 60 km area on the southern Yamal Peninsula in November 2006 (Forbes, 2008; Forbes and Stammler, 2009). The same area was affected by an even more extensive icing event a few months later in January 2007, covering an area approximately 60 × 100 km and causing the death of a large number of reindeer (Forbes and Stammler, 2009; Bartsch et al., 2010). Although the November event was considered unusual, the January event was deemed to be outside the experience of the herders interviewed.

In Fennoscandia, ice crust formation causes reindeer mortality (Moen, 2008) and calf production is reduced (Helle and Kojola, 2008). For example, during the last week of December 2007, a sudden period of warmer weather ruined good grazing pastures in northern Sweden. "*The mild period arrived suddenly, and in one week the temperature was 10 °C. Most of the snow melted and then froze again, and the ground was covered in ice. Only two weeks before we thought the guohtun (powder snow) would be good, and now it was completely inaccessible, locked away under the ice.*" (R. Stokke, pers. comm., 30 August 2008, Jokkmokk Sweden).

This was probably the same event that damaged vegetation over large areas of northern Norway and Sweden (Bokhorst et al., 2009). In contrast to the deaths of wild animals in response to extreme winter events, the impacts on semi-domesticated reindeer are mitigated by the costly process of supplementing the natural diet with food pellets.

Future warming could be associated with events that cause sudden and dramatic step changes, rather than progressive trends, in animal and plant biodiversity (see Section 4.4.3.1 for interactions between snow and vegetation).

4.4.3.3. Interactions between snow and biogeochemical cycles

Until a decade ago, Arctic ecosystems were considered inactive in terms of trace gas fluxes during the long winter period, but recent studies show that winter respiration of CO_2 represents a significant portion of the annual carbon budget (e.g., Fahnestock et al., 1998). As snow acts as an insulator and increases ground temperature, projected increases in snow depth, for example in the High Arctic, will play an important role in biogeochemical cycling during winter. For example, a record ground temperature recently occurred abruptly in the central Lena River basin, Russia, following increasing precipitation (Iijima et al., 2010).

Soil temperature and vegetation type are the main drivers of respiration rates during winter in the Arctic (Grogan and Jonasson, 2006). However, vegetation types only influence respiration in areas where the snow depth is below a certain threshold (~1 m deep). Tall vegetation that enhances snow accumulation within that threshold results in more effective thermal insulation from severe air temperatures and thereby significantly increases respiratory activity (Grogan and Jonasson, 2006). Experimental manipulation of snow depth in the Swedish sub-Arctic showed that an increase of 20 to 30 cm of snow in March (added to 74 cm and 27 cm in a birch and heath site, respectively) increased the ecosystem respiration rates by 77% and 157% compared to control plots. Later in the season, there was no significant effect of the increased snow cover (Larsen et al., 2007b). In High Arctic Canada, Nobrega and Grogan (2007) increased snow thickness from 0.3 m to 1 m, which resulted in a mean total winter efflux increase of around 60% in birch hummock tundra sites. Their results suggest that a moderate increase in snow depth can enhance winter respiration sufficiently to alter the ecosystem annual net carbon exchange from a sink to a source, resulting in a net release of CO_2 to the atmosphere, at least in the short term.

The depth of snow, in particular toward the end of the spring, will also have obvious importance for the hydrology at the onset of the growing season. This, in turn, will have strong impacts on the degree of waterlogging and, hence, the potential emissions of other important greenhouse gases, such as methane (CH_4) and nitrous oxide (N_2O); in that emissions will increase with the extent of waterlogging.

The date of snow melt is important for the growing-season accumulated uptake of carbon, as solar radiation input is high at this time compared with the period in autumn when the snowpack is formed (Callaghan, 2005). Euskirchen et al. (2006) estimated that for each day that the growing season increased across the pan-Arctic region over the period 2001 to 2100, the carbon drawdown increased by 9.5 gC/m^2 per year.

4.4.3.4. Geochemical processes in the snow cover

Analysis of the heavy metal content in snow provides useful information on aerosol composition and long-range distribution patterns of anthropogenic substances emitted into the atmosphere at lower latitudes (Colin et al., 1997; Walker et al., 2003). The effects of heavy metals are many and varied and depend on factors such as the particular metal, the organism, and the mode of uptake (Dietz et al., 1998). Their greatest toxic impacts occur in animals at the higher trophic levels where bioaccumulation through the food chain can result in high concentrations. The few available data on heavy metal concentrations in snow on ice surfaces in the Russian Arctic seas (Melnikov et al., 1994; Gordeev and Lisitzin, 2005; Shevchenko, 2006; Golubeva, 2007) mostly show no evidence of pollution, except in areas near the large industrial centers of the Kola Peninsula (Monchegorsk, Nikel) and Siberia (Norilsk) and along the Northern Sea Route. In the near future, new pollution from the oil and gas industry in the Barents and Kara Seas is possible (AMAP, 2007).

Many factors have led to a significant reduction in on-ice snow volume. These include a decrease in SCD on the Russian Arctic coast (Bulygina et al., 2009); a reduction in Arctic sea-ice extent at the end of the melt season in September (Stroeve et al., 2007); changes in precipitation patterns , particularly shifts from snowfall to rainfall; a delay in the onset of autumn ice formation; and a reduction in multiyear ice (Harbeck et al., 2008). These processes lead to changes in the size of the storage reservoirs for contaminants such as heavy metals (especially mercury), as well as persistent organic pollutants, such as PCBs (polychlorinated biphenyls) and DDT (see Chapter 11, Section 11.3). Changes in the size of the storage reservoirs, and in the processes driving these changes, could increase the delivery of heavy metals to the tundra and coastal ecosystems through changes in dry deposition and wet deposition through rainfall and fog. However, at the same time, heavy metal accumulation in snow cover on drifting ice will decrease, reducing the amounts of heavy metals transported by sea ice to areas of seasonal melting (i.e., Fram Strait and the northwestern part of the Barents Sea). This could result in a redistribution of heavy metal pollution.

Heavy metals may be deposited in greater quantities over marine areas than terrestrial areas because of the nucleating effects of sea salt, and increased rates of coastal and marine ice-fog formation and fog-related deposition (Garbarino et al., 2002). Conversion of gaseous elemental mercury to halogenated compounds following photochemical reactions in the spring may also contribute to increased levels of mercury over sea ice and water. In areas where snow is not present on the sea ice, the mercury can move more quickly into the seawater through cracks in the ice cover.

Long-term observations on the transport of air masses toward three sites in the Russian Arctic (Vinogradova and Ponomareva, 2007) suggest a reduction in the levels of anthropogenic pollutants in the aerosols over the central part of the Russian Arctic and thus a decrease in the total quantities of these pollutants deposited to the Arctic surface each year. This may result in rivers becoming a proportionately more important sink for these pollutants, if the decrease in atmospheric levels within the Arctic is due to greater deposition to rivers, snow and soil in sub-Arctic river catchments (i.e., closer to the sources). However, for lead, cadmium, and zinc the Arctic is likely to become a more effective trap, because precipitation is projected to increase (Macdonald et al., 2005).

Concentrations of gases in the Arctic atmospheric boundary layer are affected by chemical reactions on ice surfaces in snowpacks. These processes are the result of a combination of factors: snowpacks are permeable, allowing exchange between snowpack interstitial air and the atmosphere (Domine et al., 2008); snowpacks strongly forward-scatter radiation, resulting in sunlight penetration well below the surface (Simpson et al., 2002); and deposition to snowpacks concentrates many compounds. Snowpack reactions affect nitrogen oxides, halogens, ozone, organic compounds, and mercury. Mercury is a particular concern in the Arctic because of its ability to bioaccumulate and biomagnify in food webs and because it can be toxic to biota even in very small quantities. Snow plays a significant role in accumulating and redistributing mercury (Johnson et al., 2008; Steffen et al., 2008). At two sites near Barrow, Alaska, surface snow mercury concentrations increased by 15 to 30 times during a nine-day 'atmospheric mercury depletion event' (a period during which there is a sharp drop in the concentration of gaseous elemental mercury in the lower atmosphere), and much of this mercury was retained until snow melt (Johnson et al., 2008). The mercury concentrations in surface water in a small creek near the sampling sites ranged from 10 to 15 ng/L (i.e., were elevated due to the deposition event) during a ten-day peak snow runoff period (Douglas et al., 2008).

Snow contains nitrate owing to the deposition of atmospheric nitric acid and particulate nitrate. Exposure of the snowpack to ultraviolet radiation (wavelengths below ~340 nm) results in nitrate photodissociation and the formation of nitrogen dioxide (NO_2), nitrite, and hydroxyl radicals (OH) (Honrath et al., 2000; Chu and Anastasio, 2003). NO_2 formed near the ice surface can be released into the snowpack interstitial air (Boxe et al., 2003). The resulting efflux of nitrogen oxides (NO_X, i.e., NO and NO_2) from the snowpack has been measured (Jones et al., 2001; Beine et al., 2002; Honrath et al., 2002; Oncley et al., 2004). Nitrite is itself photodissociated, forming nitrogen monoxide (NO), and under sufficiently acidic conditions, can also be released in the form of nitrous acid (HNO_2) (Zhou et al., 2001; Amoroso et al., 2006). HNO_2 fluxes have also been measured (Zhou et al., 2001; Honrath et al., 2002; Beine et al., 2003, 2005, 2006; Amoroso et al., 2006) but vary widely; this is attributed to variations in snowpack pH (Beine et al., 2003), although measurement biases have also been suggested (Liao et al., 2006; Kleffmann and Wiesen, 2008). OH formed from nitrate photolysis, and in greater amounts by photolysis of hydrogen peroxide (Chu and Anastasio, 2005; France et al., 2007), is a strong oxidant of organic compounds. It is expected that OH oxidation of ubiquitous snowpack organic matter (discussed in the review by Grannas et al., 2007) occurs and contributes to observed enhancements of carbonyls and carboxylic acids in sunlit snow as well as the processing of anthropogenic organic pollutants that have deposited (Sumner and Shepson, 1999; Dibb and Arsenault, 2002; Grannas et al., 2004). Furthermore, sunlight absorption by uncharacterized organic compounds is significant and may initiate additional, as yet uncharacterized, photochemical reactions in the snowpack (Anastasio and Robles, 2007).

Ozone is destroyed in sunlit snow by a photochemically initiated process (Peterson and Honrath, 2001; Albert et al., 2002; Helmig et al., 2007) (and, at a slower rate, in the dark; Albert et al., 2002). The underlying mechanism is not yet clear. However, ice-surface reactions that release bromide ion as Br_2 or bromine monochloride (BrCl) are believed to be important sources of the active bromine (Br) that is responsible for Arctic boundary layer ozone depletion (see review by Simpson et al., 2007). It is believed that such 'bromine explosion' reactions occur to a significant extent on snowpack surfaces (Simpson et al., 2005; Piot and von Glasow, 2008). Br_2 has been observed in snowpack interstitial air (Foster et al., 2001), and it has been suggested that Br release into interstitial air may be responsible for snowpack ozone destruction even far from the ocean (Peterson and Honrath, 2001). The atmospheric impact of the snowpack ozone destruction has been observed at Summit, Greenland, during summer, where it is the apparent cause of a ~1 ppb diurnal variation in ozone concentration (Helmig et al., 2002). Owing to strong surface stability, simulated Arctic ozone concentrations are very sensitive to small variations in the poorly constrained ozone deposition velocity to the Arctic snowpack (Helmig et al., 2007), and ozone destruction within the snowpack, therefore, reduces boundary layer ozone concentrations.

The effects of air-snow surface exchange fluxes are amplified under the stable atmospheric conditions frequently present in the Arctic. Snowpack emissions appear to dominate the springtime budget of NO_X in the stable Arctic atmospheric boundary layer when relatively fresh combustion emissions are absent. Enhanced NO_X concentrations with mid-day peaks attributable to snowpack emissions have been observed at a variety of Arctic locations, with peak NO_X reaching 30 to 50 parts per thousand (ppt) (or NO reaching 10 to 15 ppt) during spring at North American locations (Ridley et al., 2000; Beine et al., 2002; Honrath and Jaffe, 1992) and 50 to 100 ppt NO_X during summer on the Greenland Ice Sheet (Yang et al., 2002). (These mixing ratios are significantly higher than those at 30 m altitude (Ridley and Orlando, 2003), reflecting the strong surface stability.) Owing to their controlling impact on boundary layer NO_X, snowpack emissions also appear to dominate atmospheric total nitrate above the Arctic snowpack (Morin et al., 2008).

4.4.4. Socio-economic processes and the role of snow cover

4.4.4.1. Water resources and hydropower

Electricity demands are rising in some areas of the Arctic, such as the Northwest Territories and Nunavut in Canada, due to increasing population and heavy industry (Furgal and Prowse, 2008). Projected changes in snow cover and duration will affect the capacity and operations of current and future hydroelectric developments and might resolve some of the rising needs. A more even distribution of water discharge will reduce the need for peak reservoir levels without generating maximum electricity, particularly if precipitation increases (Gode et al., 2007). In areas where winter precipitation is expected to increase (see Section 4.3.2.2), an increase in runoff is also expected. In Swedish sub-Arctic Suova, the increase in runoff is expected to be as much as 53% (ECHAM4OPYC3 model with IPCC A2 emissions scenario) by the end of the 21st century due to increased precipitation (Table 4.3, Gode et al., 2007).

Table 4.3. Projected percentage increase in runoff using HadAM3H and ECHAM4/OPYC3 for Suova in sub-Arctic Sweden using the A2 and B2 emissions scenarios (Gode et al., 2007).

Year	HadAM-A2	HadAM-B2	ECHAM-A2	ECHAM-B2
2011 – 2040	5.5	3.6	24.1	15.9
2071 – 2100	12	8	53	35

Projected increases in winter rainfall and increased freeze-thaw cycles are expected to lead to an enhanced winter snow melt and a decline in winter storage and, hence, a more even runoff over the year (Gode et al., 2007; Furgal and Prowse, 2008). The amount of water available for hydropower is also dependent on glacier runoff and this is projected to increase in some areas and decrease in others during the present decade (see Chapter 7). In Sweden, production from hydropower was projected using the EMPS model to simulate the potential use of power plants with increasing runoff and using knowledge on current facilities

and potential expansion, etc. The results suggested that the greatest changes in runoff are expected to occur in winter, although there was a more even distribution throughout the year (Table 4.4; Gode et al., 2007). The large variation between the results for the two GCMs used provides a challenge for economic forecasting.

Table 4.4. Changes in runoff and production calculated using the EMPS model on a seasonal basis in Sweden (Gode et al., 2007).

	Reference period 1960 – 1991			HadAM-B2 2071 – 2100			ECHAM-B2 2071 – 2100		
	Winter	Summer	Annual	Winter	Summer	Annual	Winter	Summer	Annual
Runoff (TWh)	12.5	53.5	66.0	21.5	49.9	71.3	30.4	51.2	81.6
Production (TWh)	34.2	28.1	62.3	34.8	31.5	66.3	39.4	35.1	74.5

Current hydroelectric power plant capacity and design are based on climatological and hydrological statistics, but as the climate changes, conditions for the power plants will alter. For example, in some regions reservoir capacities may need to be expanded to offset changes in runoff, both for seasonal and total annual runoff. In addition, the construction of hydropower dams needs to take into account the different future projections for ice conditions (see Chapter 6).

4.4.4.2. Infrastructure development

There has been a rapid and ongoing increase in the development of large-scale wind farms, mines, and oil and gas fields in the European Arctic. This construction entails a relatively widespread infrastructure of roads, buildings, and power lines, often within important reindeer pasture. Disturbance is a source of stress for the animals while they are within or near a development area (Vistnes et al., 2001). Avoidance of certain reindeer pastures can lead to less optimal range use, complications with herding, increased costs, and reduced production. Negative effects such as these may be exacerbated by changing snow conditions that further limit access to food (Kumpula et al., 2007).

Although cumulative impacts are likely to be important (e.g., McCarthy et al., 2005), reindeer herders in the Nenets Autonomous Okrug and the Yamal-Nenets Autonomous Okrug consider unbridled oil and gas development on their territories to be a far greater threat to their resilience than extreme weather such as winter thaw and icing events (Nuttall et al., 2005; Forbes 2008; Forbes and Stammler 2009). This contrasts with the situation elsewhere, such as Nunavut, Alaska, and Sápmi, where scientists and some indigenous politicians have sounded the alarm about climate change (Forbes and Stammler, 2009).

4.4.4.3. Transport

Rising temperatures in the Arctic, particularly decreases in very cold days, later onset of seasonal freezes, and earlier onset of seasonal thaws (as projected for many Arctic areas – see Section 4.3.2), could mean reduced costs of snow and ice control, as is the case in the mid-latitudes with seasonal snow cover. In the United States, snow and ice control accounts for about 40% of annual highway operating budgets in the northern states (NRC, 2008). Climate variables of particular relevance to the transportation sector in Alaska include the extent of sea ice, snow cover, and permafrost. In many northern states, for example, rising winter temperatures will decrease snow and ice removal costs, lessen adverse environmental impacts from the use of salt and chemicals on roads and bridges, extend the construction season, and improve the mobility and safety of passenger and freight travel through reduced winter hazards (NRC, 2008). Climate models project a shortening of the snow cover season over most of the Arctic. This will decrease the operational period of winter road networks that are important for low cost resupply of many northern communities in Canada (Furgal and Prowse, 2008) and shorten the period that snowmobiles can be used for hunting and reindeer herding. Changes in the amount and timing of the snow cover will also affect the transportation platform that lake and river ice provide in the Arctic during winter (see Chapter

6). Winter roads in Russia (*zimnik*) will be affected differently in different parts of the Russian high latitudes. In general, these roads can be used when temperatures are below -7 °C (Peschanskiy, 1967).

4.4.4.4. Forestry

Where precipitation in the boreal region is most abundant (eastern Canada, western Russia, Fennoscandia), a increase in temperature can often result in increased tree growth in healthy trees (Saxe et al., 2001) and a northward and upward extension in range. Many models project a general northward movement of the boreal forest under a warming climate, that will displace between 11% and 50% of the tundra within 100 years (Callaghan, 2005; Vygodskaya et al., 2007; Sitch et al., 2008; Wolf et al., 2008; Tchebakova et al., 2009) in a pattern similar to that which occurred during the early Holocene climatic warming (see references from Callaghan, 2005). Accelerated tree growth and range extension are important commercially for forest products, recreation, conservation, and capture of atmospheric carbon (Goodale et al., 2002). However, the moisture regime in much of the boreal region is only marginally suitable for forest growth and development. Consequently, reduced growth of boreal trees during warming was widespread over much of the past century in nearly all regions and tree species (Lloyd and Bunn, 2007). A declining trend in seasonal photosynthesis on the boreal landscape, as indicated by satellite sensors over the past 25 years, is the main trend in vast areas (Goetz et al., 2005).

New direct experimental results demonstrate that in at least some boreal evergreen conifer species, growth is largely accomplished by moisture supplied from melting of the winter snowpack (Yarie, 2008). In most of the boreal forest region, temperature increases have made the snow-accumulation season shorter and the warm season longer (see Section 4.3.1.2), so that less of the annual water budget is introduced into the ground in the spring pulse of snowmelt. Broadleaf trees can use the variable summer rains more effectively than most conifers, so this change in the seasonal type of moisture (less snow, more rain) now favors the broadleaf trees over conifers and mosses (Juday, 2009). Boreal broadleaf tree ecosystems are characterized by less storage of carbon than conifer-dominated forests owing to faster and more complete decomposition of the dead organic matter they produce (Gower et al., 2001), and this change could be an important positive feedback to climate warming. In addition, the coniferous trees are more widely used in the timber industry.

Tree rings of northern treeline trees show that even at the northernmost high-elevation treeline in western North America, 40% of trees are limited in growth by temperature-induced drought stress, which occurs at temperatures above an identifiable, consistent threshold (Wilmking et al., 2004). Tree growth patterns across the entire Northern Hemisphere at high-elevation and high-latitude northern treeline sites now show distinct evidence of 'divergence', in which formerly positive tree growth responses to increases in temperature are weaker, non-existent, or even reversed (D'Arrigo et al., 2004; Wilmking et al., 2004) because recent high temperatures are consistently above the threshold causing drought limitation of growth (Wilmking et al., 2005). Under warmer climatic conditions, the northern treeline would probably not advance uniformly into regions that were formerly too cold (as generally perceived), but would advance in a fragmented manner by occupying parts of the landscape with sufficient moisture, for example from snow accumulations.

In recent decades, conifers on productive, low-elevation sites in the interior of Alaska have been stressed to near their adaptive limits within their current North American distributions (Thompson et al., 1999; Barber et al., 2000; Juday et al., 2005). In central Alaska, spruce growth was correlated with seasonal soil-moisture deficit in unthinned control plots, but not in thinned and fertilized plots (Yarie et al., 1990), suggesting that these management treatments could partly relieve moisture stress. Across south-central Canada, the broadleaved tree aspen (*Populus tremuloides*) experienced a major collapse in productivity during the severe drought of 2001 to 2003. Extensive areas of mature aspen stands died from recent extreme weather, almost certainly because of acute drought stress, insect defoliation, and thaw–freeze events (see Section 4.4.3.1.4) (Frey et al., 2004). These developments have come as wood harvested from aspen trees has assumed greater importance as a commercial forest product in recent years because of

technological changes that allow or even require the use of its particular fiber characteristics in new products (Mackes and Lynch 2001).

Snow is an important factor in the decline of one of the most valuable timber species in North America. Alaskan yellow-cedar (*Chamaecyparis nootkatensis*) in the northernmost coastal rainforest region of British Columbia and southeast Alaska has experienced several waves of tree death over an area of more than 200 000 ha (Hennon et al., 2008). The cause of the decline is winter thaw events (see Section 4.4.3.1.5). The net effect of these winter thaw events is the loss of commercial benefits from the species for a number of centuries.

In contrast to the beneficial effects of a deep snow cover on tree growth by providing water, heavy snowfall damages forests in Finland. A modeling assessment of risk, in terms of the number of days per year when the accumulated amount of snow exceeded 20 kg/m^2, projected that compared to the baseline period 1961 to 1990, the risk of snow-induced forest damage and the amount of damaging snowfalls were predicted to decrease, particularly in the north of Finland (Kilpelainen et al., 2010).

Snow anomalies, as they interact with rising temperatures, also affect the insects that feed on boreal tree species. The European spruce engraver beetle (*Ips typographus*) takes advantage of weak and injured conifers. The beetle population increased in trees damaged by heavy storm and snow damage in the late 1960s and reached outbreak levels during a series of warm years with an acute shortage of moisture in the early 1970s (Heliövaara and Peltonen, 1999). From 1971 to 1981, severe outbreaks of *Ips typographus* in southern Norway damaged trees totaling 5 million m^3 of timber (Bakke 2006).

The North American engraver beetle (*Ips perturbatus*) is a wood-boring species that attacks already weakened trees, primarily white spruce (McCullough et al., 1998). Extensive tree injury from increased forest fires, climatic stress, and extreme snow events of the past few decades have created optimum conditions for engraver beetle outbreaks (Werner et al., 2006). Cumulative tree mortality is now heavy in many parts of Alaska (Werner et al., 2006). The complex interactions of engraver-beetle tree host, snow, and temperature were displayed during monitoring of a major outbreak in an experimental forest in central Alaska. Engraver beetle populations initially began to increase in abundant injured trees following a 1983 forest fire (Holsten, 1986). During winter 1984/85, heavy snowfall in the early winter broke branches and tops of mature spruce, and the already high engraver beetle populations increased further during spring and summer 1985 (Werner, 1993). An abnormally low snowfall followed in winter 1985/86, producing drought-like conditions in spring 1986 that weakened spruce and rendered it susceptible to a very large engraver beetle outbreak that resulted in a regional episode of high spruce mortality (Holsten, 1986). This formerly uncommon sequence of high early (November and December) snowfall followed by snow drought the next winter has become more frequent in central Alaska in recent decades. A continuation or intensification of these climate trends would reduce the evergreen conifer component of the forest. Because the conifers produce a generally more commercially valuable wood than hardwoods, and because conifer wood is less rapidly decomposed, such a shift has implications for economic return and carbon storage.

The Siberian silkmoth (*Dendrolimus superans sibiricus*) is one of the major outbreaking insect species in northern Eurasia and has a major control on the establishment and survival of Siberian conifers (Siberian fir *Abies sibirica*), Siberian pine (*Pinus sibirica*), Siberian spruce (*Picea obovata*), and Siberian larch (*Larix sibirica*) (Kharuk et al., 2003). The northern border of silkmoth outbreaks was historically represented by a growing degree-day heat sum (10 °C threshold) of 1400 to 1600 °C (Kharuk et al., 2004). During defoliating outbreaks of this insect, as many as 4.0 million hectares have been affected. The full silkmoth life cycle usually takes two years, but in warmer conditions one generation can develop in a single year, while in colder conditions up to three years may be required. Drought can induce some larvae to shift to the shorter 1-year life cycle, so that the adults of two generations emerge simultaneously, increasing the population sharply (FAO, 2009). Historically, outbreaks occurred at 8- to 11-year intervals following a few years of water shortage, and outbreak cycles are now more frequent (FAO, 2009). Within

outbreak areas, low elevations and warm, dry, steep slopes experience higher levels of tree mortality (Kharuk et al., 2007). So far, cold climate and sufficient moisture have limited outbreaks to areas generally south of 60° N (Kharuk et al., 2004). However, in the late 20th century, temperature increases and periodic low snow accumulation years facilitated the movement of the Siberian silkmoth westward in Russia, and the species now represents a serious risk for coniferous forests of Belarus, Baltic nations, and the Nordic countries (Gninenko and Orlinskii, 2002).

4.4.4.5. Natural hazards

Snow avalanches are well understood and safety measures are taken in many alpine regions, but the distribution, frequency, and impacts are less well known in the Arctic. However, climate models project warmer winters, increased snowfall, more frequent mixed precipitation, and likely increases in snowfall amounts during extreme events, which may change avalanche potential in some regions of the Arctic. A number of physics-based snowpack models exist that can be used to estimate change in avalanche potential (e.g., Guseva and Golubev, 1989; Brun et al., 1992; Lehning et al., 1999). To date, there has been no systematic assessment of change in avalanche potential over the Arctic in response to climate change as the snowpack models require very detailed site-specific input data. Lazar and Williams (2008) gave an example of the application of the SNTHERM physical snowpack model to estimate potential change in wet avalanche potential in the Aspen ski areas in the United States. Ongoing research at Svalbard aims to understand the response of mountain slope processes to a changing climate to provide a better understanding of its impacts on avalanche potential (www.skred-svalbard.no).

Slush torrents in the Arctic (Figure 4.25) are events that occur within minutes but often have return frequencies over decades and thus are difficult to document (Bull et al., 1995; Gude and Scherer, 1995). Nevertheless, they have claimed lives and are a major driver of the geomorphology of northern mountain landscapes.

4.4.4.6. Tourism and leisure

Arctic tourism is based on scenery (ice- and snow-scapes, mountains and tundra, vistas), unique Arctic wildlife (including guided hunting), and traditional indigenous cultures. Climate change could dramatically alter all of these.

Tourism in the Arctic is, at best, a marginal enterprise and is highly vulnerable to positive and negative shifts in demand. With the exception of some major cruise lines traveling to the Arctic from Europe and North America and some Alpine ski operations, it is mainly an industry of small-scale operations. With only Alaska as a possible exception, there is no 'mass tourism' in the Arctic, and the huge resorts common at lower latitudes are effectively nonexistent (Pagnan, 2003). The range and extent of tourist activity includes traditional hunting and fishing, expedition-style and destination cruising, dog sledding, and cultural and aboriginal tourism (Notzke, 1999). There is a long tourism history in the Arctic with northern mainland Norway and Svalbard featuring centrally in early Arctic travel (Viken, 2006). In addition to ship travel, there is considerable air travel and road travel, the latter more prevalent in the European Arctic where access to destinations such as North Cape, Norway, is good (Pagnan, 2003). Accurate numbers of tourists are hard to provide for the Arctic.

Arctic tourism faces three main challenges. The first is the high costs of transportation and of maintaining infrastructure in a harsh setting. The second is the short summer tourist season, in effect only a few weeks overall, which makes it heavily dependent on favorable weather conditions. The third is that Arctic tourism is essentially nature-based tourism with limited opportunity to diversify, meaning that any disruption of the natural setting or the wildlife on which the industry depends can have serious, long-term effects on the industry.

The decline in polar bear populations is at least partly due to changing snow conditions; female polar

bears make dens in snow to shelter and give birth to young during winter. Cub mortality may increase when higher temperatures lead to rains early in the breeding season, which can melt the under-snow lairs that cubs need for shelter (Smith and Harwood, 2001; Stirling and Smith, 2004). Loss of sea ice (see Chapter 9) that limits the area for snow accumulation that can be used for denning will cause further stress. The polar bear is the icon of the Arctic world. Its disappearance would be a tragic loss and would greatly diminish the polar experience (Lemelin, 2005).

Scenic attractions will change, resulting in site-specific as well as regional challenges and opportunities. Pagnan (2003) pointed out that the tourism industry in the Arctic relies on traditional perceptions of the Arctic environment and expectations about the experience that relate to ice and snow, mountains and tundra, and wildlife. What it will mean for tourists when the inaccessible becomes accessible, and the 'inhospitable' climate appears more hospitable, is not clear If tourist numbers continue to grow, related in part to changing environmental conditions, perceptions of crowding may well begin to replace perceptions of solitude. However, perceptions of dramatically rapid change in polar regions, constantly stressed by the media, may help increase public interest. Stewart (2007) and Steward et al. (2007) speculated that land-based tourism activities, such as sport hunting, eco- and nature tourism, retreat tourism, conference tourism, and winter-based tourist activities, could play a more prominent role in western Canadian Arctic communities in the future.

Snow conditions and their impacts on tourism are not well addressed in the climate change literature. Snow accumulation, duration, and consistency will have an effect on access to attractions and snow-based activities such as dog sledding, skiing, reindeer sledding, skidoo touring, and winter activities in general. Greater snow accumulation leads to access difficulties, but also to new opportunities for snow-based activities (Dawson et al., 2007). Shorter snow duration leads to seasonal challenges for some activities. Change in consistency of the snowpack, including more tundra ice versus snow leads to winter tourism challenges (such as for dog sledding, skidoo touring), and the unpredictability of snow conditions (e.g., winter thaws) could lead to negative experiences for tourists and even hazards, such as avalanches, slush torrents, and floods. Snow conditions affect ecosystems, biodiversity, and landscape amenity value. If scenic value is diminished through environmental change associated with changes in snow, then tourism will be affected.

4.4.4.7. Indigenous cultures, indigenous knowledge, and traditional land use

The Arctic environment is the setting for its indigenous peoples, containing the vital resources on which their livelihoods and cultures depend. Climate change and its consequences are of critical importance to the cultural and economic well-being of Arctic peoples (Snyder and Stonehouse, 2007). Because snow is persistent and dominant in the Arctic landscape, it plays a fundamentally important role that shapes day-to-day life, transport, and resource use (such as reindeer herding and hunting).

An extensive and detailed indigenous knowledge of snow and ice conditions and their effects is sometimes complementary to scientific knowledge. This traditional understanding is reflected in indigenous languages (Ruong, 1964; Turi, 1966; Eira, 1984; Jernsletten, 1997; Ryd, 2001; Krupnik et al., 2004; Magga, 2006; Weihe, 2006). Concepts of different types of snow range from new, light types of snow; to snow transformed by wind and weather and by grazing, digging, and trampling; to ice-related types of snow (Riseth et al., 2010). Of particular importance is the deep snow found near the coastal areas of the Finnish herding region (Helle and Kojola, 2006) and Finnmark, Norway (Tømmervik et al., 2009), which has diminished reindeer numbers in these areas. Also important are the ice layers described in Section 4.3.1.3.1.

An additional concept not yet addressed in snow and ecological research is the ground conditions during the formation of the durable snow of winter, already recorded in 1910 by Turi (Turi, 1966: 53–54; Riseth et al., 2010). Kumpula and Colpaert (2003) reinforced the importance of these conditions by maintaining that a hard snow or ice layer (bottom crust) that hampers foraging throughout winter is probably more

important than actual snow accumulation in open, high pasture areas. In contrast, mold formation occurs especially after mild, rainy autumns, when the soil does not have time to freeze before the snow falls (Eriksson, 1976; Pruitt, 1984; Turunen et al., 2009). Kumpula and Colpaert (2003) suggested that a thick snow layer on unfrozen soil in early winter occurs especially in woodland areas (that could be a future analogue of current tundra) and that it promotes mold growth that is harmful to animals and can kill calves.

Thus far, such insights have not been explored by science, but they highlight the sensitivity of the reindeer grazing system to changing snow conditions beyond the routine snow monitoring programs. As changes in climate and snow conditions render reindeer herding more insecure and less economically productive, there is likely to be a climate-induced shift from a mixed economy to a market-based economy. This would have significant cultural implications for indigenous peoples in the Arctic, even though the current general low-profit situation does not seem to limit the cultural value of herding (Forbes et al., 2006; Caballero et al., 2007; Forbes, 2008).

4.4.4.8. Human health

Human health status reflects the complex interaction of all the individual, social, cultural, nutritional, and socio-economic factors together with the environment (e.g., landscape and snow cover affect the health and well-being of Arctic populations). In the circumpolar north, climate is a major, constantly changing component of the environment. There are two categories of climate-related effects on human health: direct impacts (such as temperature and ultraviolet light), and indirect impacts (such as climate-induced changes in wildlife and zoonotic diseases; Parkinson and Butler, 2005). Other important issues include contaminant concentrations, traditional food security, community adaptation to stress, and community-based monitoring. The impact of climate change on wildlife species is critical to the diet of indigenous residents following a traditional way of life.

Societal change and modernization have many negative consequences for human health, including social and mental health problems as well as increased prevalence of chronic diseases (such as cardiovascular disease and diabetes) and alcohol abuse (Bjerregaard et al., 2004). Changes in climate, such as changes in snow conditions, are likely to add to the ongoing negative impacts of general societal change and modernization in the Arctic. The psycho-social stress is reflected in alcohol abuse, violence, and suicide, and these have all been shown to be associated with changes in lifestyle and living conditions (Curtis et al., 2005) that are affected by the impacts of changing snow conditions on traditional activities (subsistence hunting and gathering as well as recreational activities). The most important health implications of the alcohol abuse are accidents (e.g., drowning and falls), violence, and traumas (e.g., cuts, fractures, head injuries). Currently, many Arctic indigenous and other resident populations are experiencing high mortality rates from injury and suicide and higher hospitalization rates for infants with pneumonia, meningitis, and respiratory infections (Heikkinen et al., 2008; Meyer et al., 2008).

Climate change is already affecting Arctic species, including infectious disease agents, and greater changes are expected (Parkinson, 2008). The thawing of permafrost, extreme weather events (e.g., flooding due to snow melt), and storms (e.g., blizzards) may destroy infrastructure for sanitation and drinking water, leading to an increase in food- and water-borne diseases and respiratory infections (Parkinson and Butler, 2005). Higher ambient temperatures may result in an increase in some temperature-sensitive food-borne diseases, such as gastroenteritis and poisoning. Higher temperatures will benefit free-living bacteria and parasites as well as insects (Bradley et al., 2005). In addition, climate warming may drive increased dissemination of zoonotic pathogens (e.g., *Giardia*, *Toxoplasma*, and *Echinococcus* species) in water- and food-borne pathways, especially in communities using wildlife as a food source. Small changes in temperature and humidity affect the distribution and behavior of the many vector-borne diseases, as was the case with West Nile virus (Petersen and Roehring, 2001; Parkinson and Butler, 2005; see more, Emerging Infectious Diseases, 2008).

Many communities are now vulnerable to the spread of new and emerging infectious diseases, such as influenza, severe acute respiratory syndrome (SARS), and anti-microbial drug-resistant pathogens (e.g., *Staphylococcus aureus* and tuberculosis) (see more, Emerging Infectious Diseases, 2008). The release of antibacterial pharmaceuticals from human settlements may also change resistance under the cold Arctic conditions (Kallenborn et al., 2008). Climate warming and melting ice and snow may also increase the release of persistent environmental pollutants, which can compromise the immune system in animals and humans and thus increase risk of disease (Kraemer et al., 2005; Hansen et al., 2008).

Wintertime mortality and morbidity will decrease with higher environmental temperatures (Näyhä, 2005). However, extreme weather events, such as storms including blizzards and weather variability, will probably cause adverse health outcomes, especially for the elderly, those with chronic diseases, and children. The physiological adaptive capacity may decrease due to urbanization and an ageing population (Hassi et al., 2005; Mäkinen 2007). Low temperature and low humidity have been associated with increased occurrence of respiratory tract infections, with a decrease in temperature and humidity preceding the onset of the infections (Mäkinen, 2007).

Environmental temperature is closely associated with population mortality (Näyhä, 2005); however, in Yakutsk, Siberia, no increase in mortality has been observed even at temperatures of -48 °C (Donaldson et al., 1998). Seasonal patterns of death from suicide are well documented and have been attributed to climatic factors such as solar radiation and ambient temperature. A recent study from Finland showed that winters with low levels of solar radiation may increase the risk of suicide (Ruuhela et al., 2009). Not only cold, but also air contaminants (e.g., particulates) can increase mortality and morbidity; however, the interaction of air pollution with temperature-related mortality is not understood and needs further research.

Indigenous peoples in the Arctic are aware that climate change is occurring. Observed impacts include a significant thinning of sea- and freshwater ice, a shortening of the winter ice season, reduction in snow cover, changes in the distribution of wildlife and plant species, thawing permafrost, and increased coastal erosion of some shorelines (see Chapter 5). The predicted impacts on the environment, regional economies, and people are far reaching. The reported increase in unusual weather patterns and storm events has significant impacts on travel and hunting and fishing safety (Krupnik, 2007). For Inuit communities, sea-ice travel is critical for accessing wildlife resources and traveling between communities during winter months (Laidler et al., 2009). The implications of these changes on food security and potential implications on nutritional health among these populations, which receive significant energy and nutrient contributions to their total diet from these traditional/local foods is only now being investigated. In fact, several focused research projects have been initiated with the communities involved in the present assessment and others in these regions. For example, work on climate and water quality, hunting behavior, women's health, and emerging and chronic diseases in the Arctic are currently underway (Furgal and Seguin, 2006).

For the Inuit, disease is a consequence of transgression against the social or spiritual orders (Richmond and Ross, 2009). Inuit culture places emphasis on social behavior and its relationships with disease. At the same time, 'common sense empiricism' coexists within the Inuit belief system; snow blindness, for example, is common sense pathology and does not require a metaphysical explanation. Snow blindness has been a major problem in the circumpolar area for a long time (Sköld and Axelsson, 2008). While biomedicine is better informed about the workings of the body, traditional Inuit understanding of the relations between person, family, social group, community, and the environment represents a global view of health that is not shared by biomedicine (Bowd, 2005).

Reindeer herding among the Sámi includes many hazardous tasks in potentially dangerous environments, especially during the gathering of the reindeer for migration or slaughter. During these periods the herders use vehicles (i.e., motorcycles, snowmobiles, helicopters, airplanes, boats) to gather the reindeer, and the work is often executed during long working hours in a harsh climate and terrain. For example, most reindeer-herding men spend more than 800 hours per year on snowmobiles (Daerga et al., 2004). The

increasing number of work-related fatal accidents among reindeer herders is probably also related to an increasing pressure from the Scandinavian societies to develop profitable reindeer herding companies (Hassler et al., 2004; Tynes and Haldorsen, 2007; Soininen and Pukkala, 2008). This has resulted in external socio-economic pressure and competition between the family companies within the Sámi communities that have, in turn, forced the enterprises to make costly investments in vehicles to save time and expense on personnel. Of the cause-specific mortality, men showed an increased risk of dying from vehicle accidents, snowmobile accidents, drowning, poisoning, and 'other causes'. The number of deaths caused by snowmobiles and terrain vehicles tripled between 1961 and 2000 (Hassler, 2005). Reductions in snow cover would eventually lead to less use of snowmobiles probably because of increased open terrain unsuitable for vehicles and, perhaps, a shift away from reindeer herding in a warmer Arctic. This would result in fewer accidents. However, the psychological results might offset this, and adaptation during the transition period of changing snow conditions could be particularly difficult.

4.5. Suggested strategies for local and regional adaptation

- The main drivers of change in Arctic snow cover come from outside the Arctic, and the ability for Arctic peoples and other Arctic residents to mitigate these changes is very small.
- Adaptation is now required. While adaptation has been a continuous process in the past, the current degree of change, together with new constraints on adaptation, challenge adapative capacity.
- There is a need to use existing information effectively, but adaptation also requires better tools from the research community, such as high-resolution models of snow changes, and better communication between researchers and stakeholders.

The current degree and rate of change in snow conditions presents problems for Arctic peoples who cannot migrate as freely as previously (Chapin et al., 2004), and for many Arctic species that characteristically grow slowly, have relatively few offspring, and reproduce late in life, and thus cannot adapt quickly (Callaghan, 2005; Loeng et al., 2005). The impact that Arctic communities can have on mitigating likely future changes in snow is likely to be insignificant. As mitigation globally is likely to be a slow process, Arctic communities must adapt.

4.5.1. Natural adaptations

Many Arctic species have already developed adaptations to snow cover (see Sections 4.4.3.1 and 4.4.3.2 and Callaghan, 2005). What is unclear is whether and how they might adapt to changing snow conditions. In the short term, animal behavior (such as the timing of the start of migration) might adapt to changing snow-free periods on the tundra and the timing of plant development (phenology) might adapt to changes in the timing of spring snow thaw, first by plasticity already existing in the plants' characteristics and later by the selection of particularly well-adapted individuals. Studies show that the onset of development and behavior patterns in animals, plants, and migratory birds have changed by up to six weeks in response to earlier snow melt (Høye et al., 2007). However, both in animals and plants, the rate of mutation and true genetic adaptation to changing snow conditions is likely to be slower than the changes in snow regime. This will lead to mismatch between the distribution of species and the climate (snow regime) envelopes to which they are adapted. Examples include animals with fur or feathers that remain white when the snow has melted and which thus become more vulnerable to predation, animals that need a snow cover for nesting or denning (e.g., lemmings and polar bears), plants that require moisture from deep snow cover (see Sections 4.4.3.1 and 4.4.3.2), and plants that exist only because there is snow cover (Hodson et al., 2008) or because a persistent snow cover excludes competitors (snow bed plants). There is also a perception that earlier snow thaw will expose plants at sensitive growth stages to enhanced levels of springtime UV-B radiation. The net result of the mismatch between rapidly changing snow conditions and natural adaptations is that many Arctic species will be replaced by competitors from southern areas as they migrate northward (such as the red fox displacing the Arctic fox) (Tannerfeldt et al., 2002).

Since the publication of the benchmark knowledge in the Arctic Climate Impact Assessment (Callaghan, 2005), two perspectives on natural adaptation have developed. First, there is increasing evidence of the importance of extreme warming events in winter on the snowpack and ecosystems (see Box 4.2 and Sections 4.4.2.1 and 4.4.2.2). These events will select for existing species and genotypes that are tolerant of such extreme conditions, while those that cannot adapt will open the ecosystem to invaders from the south that are pre-adapted. Second, there is a recognition (Botkin et al., 2007; cf. Callaghan et al., 1992) that a changing climate will create new genetic variation and new adaptations to the changing conditions. Mechanisms underpinning this process include increased sexual reproduction by plants in longer, warmer growing seasons; new mixes of species as southern species move northward; and isolation of currently large populations as they become fragmented by hydrological changes on the landscape and as rising sea level create archipelagos.

Box 4.2. Changes in snow stratigraphy, impacts on reindeer, and adaptation by herders

The Sámi have a special tool (the *goaivo-soabbi*) for digging through snow to assess density and ice layers as well as vegetation condition for reindeer foraging (Figure 4.26). The stratigraphy of the snowpack is a record of snow events, temperature fluctuations, wind-compacting actions, and compaction events by reindeer. Ice layers in the upper part of the profile are less detrimental to reindeer than bottom crusts. The Sámi use various management methods, which they can apply to adapt to future changes in stratigraphy (Roturier and Roue, 2009). When ice layers are so hard that weaker reindeer cannot penetrate through the snow, the stronger animals are used to break the ice. Under even more difficult conditions, the herd must be free to search for grazing over a larger area, be moved to another area (Bartsch et al., 2010), or be given supplementary feed; if necessary, weak animals are slaughtered. In contrast, reindeer are allowed to choose their grazing freely on pastures with soft snow (Riseth et al., 2010). Management practices such as these will need to be employed more often or even supplemented by new activities as extreme events are expected to increase (Moen, 2008; Roturier and Roue, 2009).

4.5.2. Adaptations in infrastructure maintenance and development

Arctic residents need information on present and projected snow conditions to be able to adapt the built environment and urban lifestyles to the changing climate. This information includes projections of maximum snow accumulations for snow load calculations, water supply, flood management, avalanche risk, and snow clearing for permafrost stabilization. For example, snow accumulation from blowing snow and snow clearing around the periphery of raised gravel landing strips can have a major impact on the thermal regime and promote permafrost degradation that can lead to subsidence and cracking of runway edges (Allard et al., 2007).

Information that helps mitigate hazards is also required. Avalanche is a well-documented hazard in Canadian and other Arctic communities (Stethem et al., 2003), and there is potential for increased risk in the future in response to the projected increased precipitation over northern latitudes (Christensen et al., 2007). The Québec Government commissioned an avalanche hazard assessment for several communities in Nunavik in 2000 (Lied and Domaas, 2000) in response to an avalanche in the village of Kangiqsualujjuaqon on 1 January 1999 that resulted in nine deaths and 25 injuries. The study recommended the relocation of some houses and the investment of CAD 5.6 million in the construction of defense structures to mitigate the hazard.

Changes in snowpack amount and properties linked to climate change may also require revision of snow load calculations for buildings. Maximum snow accumulation is projected to increase over large areas of the Arctic (Figure 4.15) and there is potential for increased frequency of extreme precipitation years (Christensen et al., 2007). In some countries, such as Canada, snow loads are estimated from historical maximum snow depth data with a regional mean snow density taken from historical data (Newark et al., 1989). This approach ignores interannual variability in snowpack density and assumes that snowpack density does not change over time, which may not be the case under a rapidly warming climate. The

severe snow winter of 2007/08 caused a number of structural failures and several deaths in Québec and prompted a process to review and update ground-snow load calculations for Canada. Similar reviews are being undertaken in other countries in response to recent building failures and concerns that climate change may increase the snow-load hazard (e.g., Strasser, 2008).

4.5.3. Adaptations in land use and resource use

4.5.3.1. Reindeer herding

Reindeer are a major source of food in the Arctic and reindeer herding is a low-intensive land use and a low-profit industry. The changing climate, variable snow regimes, icing over of ranges, and insect harassment (Moen, 2008; Bartsch et al., 2010) are difficult to control. However, reindeer movement over large areas for grazing, usually with seasonal shifts, is a natural adaptation to the Arctic landscape and climate (ACIA, 2005; Tyler et al., 2007), with summer grazing in coastal areas and winter grazing in the taiga areas. The choice of strategy will depend on a number of factors (such as season, snow type, temperature, landscape, and the physical conditions of the animals), and will be influenced by previous experience of unusual weather conditions such as ice layers. Nevertheless, many of the problems in reindeer herding can be traced back to disturbance problems and problems with pastural balance, for example, related to property rights and the general development of infrastructure. National borders, border closure, and bilateral conventions also have restricted ability to adapt. In the past 30 to 40 years, long-range migration patterns have been restricted (Riseth, 2003; Riseth and Oksanen, 2007) in Fennoscandia and in the Yamal-Nenets area, resulting in unused and underused grazing areas (Caballero et al., 2007; Forbes and Stammler, 2009).

Reindeer response to the variable onset of seasons will initially involve changes in local migration patterns. As the snow season duration decreases, they will migrate earlier to the summer grazing grounds and stay longer on summer and autumn grazing areas. Heat and insect stress are relieved by the reindeer moving to higher areas with more wind and more snow patches; a reduction in snow beds in warmer summers will make this more difficult. In a longer perspective, reindeer husbandry needs to increase its resilience by maintaining a choice of grazing sites. Traditionally, sustainability was obtained by the use of long migration patterns. However, as noted above, this capacity has already been severely limited, while other, competing land uses further restrict the flexible use of the landscape for reindeer grazing (Moen, 2008).

Primary production in the form of graminoid availability and quality may be easier to control. Through improving preferred range quality and availability, reindeer condition will improve (Eilertsen et al., 2000). Reindeer in better condition may then tolerate stochastic and difficult climatic events, such as deep snow (that requires more energy to move over and dig through for food) or ice layers in the snow (that prevent access to food below and limit the smelling cue for the reindeer to detect food) that compromise their survival and production. In addition, varying degrees of avoidance and adaptation may mitigate some of the impacts of a changing climate and snow cover.

From the 1980s onward, complementary forage resources (food pellets) have become increasingly important (Kumpula, 2001) as an adaptive strategy to compensate for forage deficit due to restricted migrations. For example, reindeer management in Norway has used some complementary forage resources and supplementary feeding under difficult winter conditions (Åhman et al., 2002).

Reindeer herding is not only affected by changes in the snow and ice structure, but also by other factors. Adaptation cannot be explicitly focused on changes in snow conditions; it must include other local factors, and there must be recognition of indigenous knowledge systems (Tyler et al., 2007). Indigenous peoples have successfully occupied the Arctic due to their adaptive capacity (in social, economic, and cultural practices) to adjust to climate variation and change (Nuttall et al., 2005).

4.5.3.2. Conservation

Extreme events related to snow cover dynamics can have profound effects on Arctic animals and plants (Sections 4.4.2.1 and 4.4.2.2). Conservation in the Arctic is currently based on the designation and control of protected areas, of which there are many (CAFF, 1994, 1996). However, this conservation management only moderates direct impacts of humans and does not protect species, ecosystems, or habitats from climate change, such as the extreme events of winter snow thaw and icing. New paradigms are required (Usher et al., 2005) and future plans must incorporate flexibility in anticipation that changes in climate, including snow conditions, are likely to continue. For example, rather than designing static nature reserves on a landscape that is frozen in place, there should be an anticipation that disturbance will occur and climate will change (Chapin et al., 2004; Forbes et al., 2009). In addition, species may be moved further northward to a more appropriate climate envelope or fed supplementary material during crises such as rain-on-snow events. Harvest regulations are also a means of conservation management; however, regulations are often rigid and difficult to change in anticipation of changing environmental conditions. An adaptive management approach is required, which in many cases may incorporate the establishment of community-based monitoring. Major action will be required soon if populations of species such as Peary's reindeer are to be saved.

There is a need for innovation in conservation. Interactions between reindeer (*Rangifer tarandus tarandus*) and domestic sheep (*Ovis aries*) have been studied in Norway using controlled experiments and field observations (Eilertsen et al., 2000). Other research has shown that high biodiversity in northern coastal farmland can be maintained by reindeer when the farmland is abandoned, while also improving the health of the reindeer (Eilertsen et al., 2000). Biodiversity hot spots of Arctic-alpine plants in the Scandinavian high mountains that depend on disturbance from grazing are enhanced by reindeer trampling and grazing (Caballero et al., 2007). Thus, the traditional role of reindeer herding could be reinterpreted as a stewardship of the landscape by using reindeer to retard the advance of shrubs and trees and the disappearance of valued flora and fauna.

4.5.3.3. Tourism

The key challenges of climate change for the tourism industry in the Arctic relate to infrastructure, access, attractions, and cultural identity. Opportunities exist for new activities, replacement, and diversification to moderate the negative and benefit the positive impacts of climate change. Vulnerable communities are those whose local or business conditions do not currently demonstrate the capacity to change or support the flexibility to respond to change. Small tourist operations focused on a single Arctic activity, such as dog sledding, may be in jeopardy over the long term and may need to diversify.

Tourism may shift northward, as cooler regions enjoy warmer summers, while warmer regions like southern Europe suffer increased heat wave frequency, reduced water availability, and poor snow conditions in the alpine resorts. For example, Davos in the Swiss Alps experienced a decrease in snow depth of 12% between the 1960s and 1990s (Beniston et al., 2003). One study (Hamilton et al., 2005) projected that both Canada and Russia would have a 30% increase in tourists with only 1 °C of warming (Stern, 2006). Thus, the market for certain types of tourism (e.g., cruises, whale watching) could prosper in a more 'benevolent' climate. Although currently limited to a few countries, eco- and nature tourism might be expanded, perhaps by offering a new type of product from that traditionally offered. The trend of increased tourism will provide an opportunity for Arctic entrepreneurs who can adapt to the changing conditions. However, the tourism industry is not yet ready to adapt and respond. The pace of social response in the Arctic is slow and there is a limited pool of people – particularly trained individuals – to fill any new jobs. Furthermore, if there is an increase in tourism, there will need to be a development of conservation management regulations to reduce disturbance impacts, such as those ongoing in Svalbard owing to the significant increase in tourism over the past decade (Madsen et al., 2009).

4.5.3.4. Hunting and fishing

Hunting activities are particularly sensitive to snow and ice conditions (Ford and Furgal, 2009). Many indigenous groups and other Arctic residents, particularly hunters and fishers, depend on their ability to predict animal behavior in relation to nature and weather conditions. The changing climate, together with rapid economic change, modernization, and alterations in the food supply, is already affecting Arctic communities (Bjerregaard et al., 2004; Curtis et al., 2005).

4.5.3.5. Agriculture and forestry

The presence of an ice crust with a thickness greater than 20 mm during a single five-day period is considered a dangerous event for winter crops, and when it occurs it generally results in the need for replanting. The decrease in dangerous events for winter crops in a major agricultural region of the Russian Federation (Bulygina et al., 2010a) would suggest that the yield of existing crops would improve significantly and there would be fewer problems for new areas of crops as they extend northward.

4.5.4. Provision of knowledge (scientific and traditional) for adaptation

The provision of relevant knowledge for adapting to a changing snow climate in the Arctic first requires an identification of vulnerabilities and sensitivities to a changing snow regime in order to determine what snow information is required. This is no small task as there is a wide range of snow-related sensitivities in Arctic geomorphological, ecological, and human systems.

Arctic residents rely on snow for water, transportation, food storage, and for building emergency shelters (UNEP, 2007). A reduced snow cover is associated with many community-level impacts (Figure 4.27), while the potential for increased snow accumulation in some regions will require a reassessment of snow loads and the risk of avalanches and floods. Hunting activities are particularly sensitive to snow and ice conditions (Ford and Furgal, 2009) and knowledge of changes in the timing and amount of snow are important in this respect.

There is a need for better monitoring and forecasting of snow conditions and better projections of future snow conditions. A particular challenge will be to explain transient non-linear changes in phenomena, such as winter thaws and rain-on-snow events, and how these will differ in mountain regions where the sign of snow cover change (more snow or less snow) may alter with elevation.

The provision of relevant information from scientists requires the active involvement and collaboration of community members and local, regional, and national organizations that use the information for decision-making and policy development (Pearce et al., 2009). Recent research has shown that Arctic communities can exhibit a high degree of adaptability to change and can be quick to take advantage of innovation and new opportunities (Ford and Furgal, 2009) if appropriate economic measures are in place. Consequently, greater engagement is needed between the scientific community and Arctic residents in order to incorporate this traditional knowledge in activities such as, for example, interpretation of satellite-derived snow cover products; development of community-based snow monitoring programs; and attempts to better understand the impacts of uncertainties in snow information on local decision-making. The ice-edge monitoring product developed by Noetix Research, Inc., for Polar View (www.noetix.ca/floeedge) is a good example of the first, and involved extensive consultations with local hunters and elders to provide Inuktitut terminology for ice features visible on SAR imagery. In northern Fennoscandia, a comparison of indigenous knowledge on snow conditions, data from snow physics measurements, and results from experiments that simulate extreme events led to a better understanding of processes relevant to reindeer herders (Bokhorst et al., 2008, 2009; Riseth et al., 2010).

Collaboration between scientists and stakeholders will need to improve if scientists are to provide more-relevant projections of change in snow conditions. This would require that scientists evaluate and validate snow information from climate models in conjunction with the user communities in order to establish

which variables are important, the accuracy required, and the impact of uncertainties on decision-making, as well as to develop best practice guidelines for the development of snow scenarios for the Arctic. A fundamental issue is the geospatial scaling of information; adaptation is a local process and the provision of local data, such as climate downscaling (see Section 4.3.2.3), will become increasingly important.

4.6. Uncertainties, gaps, and recommendations

- Assessing current changes in snow conditions and projecting future changes and their impacts are limited by availability of standardized pan-Arctic datasets with high-resolution observations, uncertainties in remote sensing technology, and difficulty in applying highly detailed small-scale models to the pan-Arctic scale.
- A new generation of international networking is required to monitor snow and to provide data for parameterization and validation of models.
- The quality and quantity of information from the wide array of tools and techniques available to generate snow-related information for decision-makers is expected to increase over time.

4.6.1. Major issues

4.6.1.1. Resolving key uncertainties and an agenda for key research priorities

Snow cover characteristics in the Arctic are among the most rapidly changing variables associated with ongoing climatic change. *It is critical to invest time and resources in intercomparison and blending of the various types of information from existing and projected snow datasets within a context of social relevance.* Each discipline and science component – ground measurements, remote sensing, the impact research community (ecologists, hydrologists, engineers, foresters, reindeer herders, etc), economists, and adaptation planners and decision makers – have their own agendas and processes for furthering their research.

However, communication among these communities is generally poor, and there is often a mismatch in information exchange; for example, the scaling and parameters measured by ground-based monitoring and remote sensing of snow may not provide the information most relevant to ecologists, indigenous peoples, and planners. A particular mismatch is that of spatial scaling (in that adaptation requires local information in high detail) and temporal scaling (in that short-duration extreme events cannot be recorded and passed on to the impact community to allow real-time observations of impacts). Within the process as a whole, uncertainty increases along the sequence: recording and projecting snow changes → recording and projecting impacts → calculating the economic and societal consequences of the impacts → developing adaptation strategies. A particular uncertainty involves finding relevant socio-economic assessments of impacts as these are often too numerous, too local in relevance, and largely inaccessible to the wider community.

In response to the lack of a system approach, ***we recommend*** *the establishment of formal networking (e.g., similar to the Canadian ArcticNet philosophy) of all the relevant players to exchange information and to ensure that more relevance is gained in future data gathering, interpretation, and projection and that currently inaccessible data be made available. Such a network should strive to secure funding and appropriate facilities for the network components from a stronger basis than any single component. The developing process of SAON (Sustained Arctic Observing Network) could potentially fulfill many of these needs.*

4.6.1.2. Improving data acquisition, availability, and a generation of new products

There is a need to secure input information, including traditional and scientific knowledge, for snow modeling projections, and each societal service related to snow cover. The current assessment has, by necessity, drawn on various types of information from numerous archives that are often not linked or

compatible. ***We recommend*** *that research stations in the Arctic develop a holistic hydroclimatological research and monitoring capacity alongside their present disciplinary foci.* ***We also recommend*** *that operational snow-related data streams be channeled to a single point of contact. One of the two World Data Centers for Meteorology (in Obninsk, Russia, www.meteo.ru/mcd or in Asheville, North Carolina, www.ncdc.noaa.gov) or the World Data Center for Cryosphere Information (U.S. National Snow and Ice Data Center in Boulder, Colorado; http://nsidc.org) might be such a point of contact.*

In addition, ***we recommend*** *the development of an international, integrated, historical snow-related* in situ *dataset. This dataset should include all cold season* in situ *information for the period of instrumental observations (in North America after the Second World War and in Northern Eurasia for the past 100 years). To achieve this goal,* ***we recommend*** *that an international project is sponsored by the meteorological and hydrological services of the Nordic countries, Russia, the United States, and Canada. The project can be organized under the administrative auspices of the Arctic Council and under the scientific guidance of the World Climate Research Programme, via the Climate and Cryosphere (CliC) Project.*

4.6.1.3. Facilitating adaptation to changes in snow cover

Estimating the magnitude of change and its impacts as well as the requirement for models to be generally spatially applicable has focused attention on the large scale. However, adaptation is a local process responding to local changes in snow and its impacts (Section 4.3.2.3). ***We recommend*** *that there should be an increased focus on modeling and observation of changes in snow and their impacts on the local scale. Data should be made available to authorities that wish to plan adaptation strategies. Downscaling of climate, snow, and impact models is fundamental in this respect. In addition, there is a need to stimulate innovation in order to diversify resources and their usage.*

4.6.2. Detailed issues

4.6.2.1. Measurements of snowfall and snow cover parameters

- **Accurate Precipitation Information**. The provision of accurate precipitation information for the Arctic (Section 4.3.1.1) remains an ongoing challenge as the observations are prone to significant random errors and biases (Sevruk, 1982; Goodison et al., 1998) and automation of observations has created new challenges for maintaining the integrity and quality of the observations. A number of procedures have been proposed to correct for biases in frozen precipitation measurements (see Mekis and Hogg, 1999; Yang, 1999; Yang et al., 1999; Bogdanova et al., 2002a,b; Aleksandrov et al., 2005; Sugiura et al., 2006), but the corrected values still carry large random errors at the daily timescales required for many applications. ***We recommend*** *using the combination of snowfall, snow cover, and streamflow data over large Arctic river basins (e.g., Bulygina et al., 2007; Shiklomanov et al., 2007) to provide a useful basis for examining the ability to close the water budget and identify possible systematic errors in precipitation datasets.*

- **Development of pan-Arctic snow and related climate datasets**. At present, there is no pan-Arctic dataset of *in situ* snow (e.g., SWE) and climate data for understanding large snow cover changes and their impacts on regional hydrology and ecosystem functions. SWE and snow density are measured throughout the Arctic as part of SWE monitoring activities in Russia, Canada, Alaska, and Scandinavia (Brown and Armstrong, 2008), but there are large data gaps and there has been a general decrease in the number of measuring sites over time. Data are particularly sparse for the tundra region (see Appendix 4.1). ***We recommend*** *the production of gridded data for model applications, validation of remote sensing snow data and products, and the development of corresponding pan-Arctic datasets of observed snow cover changes from traditional knowledge.* This will require taking account of sources of inhomogeneity.

- **Homogeneity of time series**. There are known sources of error from changes in procedure, instrumentation, or siting of equipment and gaps in observations for snow and snow-related datasets such as solid precipitation (Section 4.3.1.1). For example, procedural changes were made in the Russian snow depth and course measurement programs in 1950 and 1966 that may have created inconsistencies in the data (Meshcherskaya et al., 1982; Krenke, 1998, updated 2004). In North America, automation at some Canadian meteorological stations (Goodison and Louie, 1986) resulted in the loss of snowfall measurements and required new quality control routines for snow depth measurements obtained from ultrasound sensors. In addition, the daily snow depth observing network over northern Canada is biased to coastal locations, and these sites can have quite a different climate to inland locations. ***We recommend*** *the development of approaches to fill data gaps and take account of biases in the observing networks.*

- **Ongoing evaluation and synthesis of satellite snow cover data**. Monitoring snow cover over the Arctic region (Section 4.3.1.2) is a major challenge for many reasons, including strong local controls on snow cover, frequent cloud cover, large gaps and biases in surface observing networks, and confusion of lake ice and snow cover during the melt season (see Appendix 4.1). The NOAA weekly snow cover product (Robinson et al., 1993) has been the most referenced source of information on variation and trends in Arctic snow cover extent since the late 1960s. However, this product includes changes in the volume and resolution of satellite information over time, as well as differences in the way patchy snow was mapped by analysts. This has resulted in inconsistencies in the mapping of snow that are particularly apparent over the Arctic (Brown et al., 2007). ***We recommend*** *the use of a multi-dataset approach for monitoring spring snow cover variability over the Arctic, as Brown et al. (2010) found this provided more robust information along with an estimate of the error in the observations. The ongoing integration and synthesis of Arctic snow cover information from multiple sources would fit well with the role of the new Global Cryosphere Watch proposed by the World Meteorological Organization (Goodison, 2009) and responds to recommendation R3.5 of IGOS-P (Key et al., 2007).*

- **Improved microwave measurements**. Satellite observations using the microwave sensor have been routinely used to map and monitor snow cover since the late 1970s (e.g., Grody and Basist, 1996; Armstrong and Brodzik, 2001; Derksen et al., 2003; Kelly et al., 2003; Appendix 4.1). The ability of microwave sensors to 'see' snow through most clouds and at night is particularly attractive for monitoring snow cover over high-latitude areas and passive microwave has proved useful for monitoring snow melt onset and snow-off dates (Takala et al., 2009; Tedesco et al., 2009) due to its sensitivity to the presence of liquid water in the snowpack. However, there are some special challenges in applying passive microwave algorithms over the Arctic region, including shallow snowpacks, extensive depth hoar, ice layers within the snowpack, and the effect of frozen water bodies. ***We recommend*** *the continued development and application of new boreal and tundra algorithms specific to the Arctic region (e.g., Rees et al., 2006; Derksen et al., 2008a, 2009; Lemmetyinen et al., 2009; Rees et al., 2010), which may overcome some of the limitations of previous global SWE algorithms. The development and testing of high-latitude SWE algorithms can be done in conjunction with intensive field campaigns at existing Arctic super-sites, as outlined in recommendation 3.4 of IGOS-P (Key et al., 2007).*

4.6.2.2. Traditional Ecological Knowledge

- **Lack of measures of uncertainty**. Science has established measures of uncertainty (e.g., statistical analyses based on probability), but the traditional ecological knowledge (TEK) used throughout this assessment does not. Some observations have high certainty, and survival has depended on them. However, other observations may be biased by particular observers. Currently, there appears to be no way of scaling degrees of certainty. ***We recommend*** *that measures of uncertainty be developed and applied.*

- **Lack of scaling perspectives**. TEK provides information at local scales and over varying time periods. Integrating across numerous local observations collected at different time intervals is a major challenge. ***We recommend*** *improved networking of community monitoring programs and centers or holders of TEK to develop wider spatial-scale understanding of indigenous perspectives of changing snow conditions and their impacts without losing the specific variability at individual locations.*

- **Lack of appropriate knowledge**. TEK has been accumulated and passed down over generations and is therefore, based on observations of past environmental conditions. However, new developments in the climate system are causing the indigenous peoples to be less able to predict weather and resource availability. ***We recommend*** *better and urgent recording of TEK and the development of a forum to discuss, develop, and record the success or otherwise of newly developing knowledge and skills applied during a changing environment.*

- **Loss of TEK**. Use of modern infrastructure and equipment (snowmobiles, motor boats, mobile phones, helicopters) together with changes in the mobility and social structure of indigenous populations has resulted in some loss of TEK that could be relevant to understanding changing snow conditions and their implications. ***We recommend,*** *as above, the urgent recording of TEK before it is lost.*

- **Comparing science and TEK**. Natural science, in contrast to TEK, has the general goal to be objective with all its results promptly validated. TEK does not have this goal. It can however, explain the practical importance of changes in snow conditions that natural science measures. This is important socio-economic information that is missed in scientific databases. ***We recommend*** *that combined methods be developed to use both scientific (instrumental) and traditional ecological knowledge more comprehensively to describe the state of snow cower, its changes, and its impact on environmental services in the Arctic (e.g., Bartsch et al., 2010).*

- **Translating snow terminology**. Indigenous peoples and scientists have their own, often sophisticated, terminology for a vast range of snow conditions. While there are dictionaries that linguistically translate some of the indigenous terms, ***we recommend*** *that an overarching standard terminology be developed, with definitions that can be used by all users and providers of snow information. In addition,* ***we recommend*** *that the terms be reassessed and that correlations (transform functions) are established among these definitions when and where it is possible. Only then will it be possible to secure input information (including traditional and scientific knowledge) for snow modeling projections and each societal service related to snow cover.*

4.6.2.3. Snow modeling and snow cover change scenarios

- **Limitations in general circulation models**. GCMs are unable to simulate the observed rate of decrease in spring snow cover over northern hemisphere high latitudes over the latter half of the 20th century (Brown and Mote, 2009; Section 4.3.2.1). ***We recommend*** *that resolving this inability be an urgent priority. The ability of models to simulate the snow melt process also needs further investigation within the context of Arctic hydrology. This should result in evaluations of feedbacks between the timing of snow melt and broader changes in terrestrial ecosystems.*

- **Scaling issues**. The scaling-up and incorporation of key Arctic snow processes to the 10–50 km resolution of regional climate models and the 100–200 km resolution of GCMs is an ongoing challenge (e.g., Déry et al., 2004). Some progress has been made in recent years to model snow and snow processes in Arctic environments (Brun et al., 2008; Section 4.3.2.3), and with sufficient good quality input data, the current generation of high-resolution Arctic snow process models such as CRHM (Pomeroy et al., 2007) and SnowTran-3D (Liston et al., 2007) can provide realistic simulations of the spatial, vertical, and temporal evolution of snowpack properties. ***We recommend*** *the continued and accelerated development and application of high-resolution Arctic snow process models.*

- **Models of snow-vegetation interactions**. Snow-vegetation interactions are numerous and complex as they depend both on changing snow conditions and on independent changes in vegetation (Section 4.4.3.1). Results from the recent SnowMIP2 comparison of snow model vegetation schemes (Rutter et al., 2009) revealed that the parameterization of key processes, such as snow unloading, are either nonexistent or based on site-specific data. In addition, the intercomparison highlighted the large uncertainties that exist in the variables required to drive distributed snow process models in re-analyses and climate models, in particular the timing, amount, and phase of precipitation; incoming longwave radiation; and cloud cover. ***We recommend*** *that improved datasets be collected for parameterization of key processes in snow vegetation models.*

- **Sublimation**. A potentially important, but often overlooked (and hence under-represented topic in this assessment) process is the sublimation of snow, especially when enhanced by blowing snow. Sublimation can be a key part of the hydrological cycle locally and regionally (Pomeroy and Li, 2000), and up to 40% of snowfall was estimated to sublimate near Tiksi, Russia (Hirashima et al., 2004). Nonetheless, climate models do not include the enhancement of sublimation by blowing snow, and some models do not include even the direct sublimation of snow from the surface. ***We recommend*** *ongoing efforts to parameterize and up-scale blowing snow sublimation processes for inclusion in climate and hydrological models.*

- **Climate model snow-cover projections**. The uncertainties in climate model snow-cover projections (Section 4.3.2.2) due to natural variability, climate model formulation, and emission scenarios need to be evaluated. ***We recommend*** *that the procedures used by Rowell (2006) for applied information, such as winter thaw events, snow season start and end dates, maximum SWE, date of maximum SWE, snow melt runoff timing and amount are used to evaluate the uncertainties. This information should then be provided to the user community, so that they are aware of the magnitude and relative contribution of the various sources of uncertainty in snow cover parameters.*

4.6.2.4. Additional issues

- **Current impacts of changing snow conditions**. Observational power in the Arctic is relatively low, and the pathways of information flow from observing particular snow conditions to identifying relevant impacts (implicit throughout Section 4.4) are insecure and often based on correlative rather than causal analyses. ***We recommend*** *capacity building to improve observational power, development of better communication pathways between the observation and impacts on communities and between the observation community and researchers who can deploy experimental methods to elucidate causes of observed changes in snow conditions.*

- **Projected impacts of changing snow conditions**. Projected impacts (Section 4.4) are determined mainly through either modeling impacts based on projections of future snow conditions (e.g., hydropower) or experiments that simulate various aspects of projected changes in snow conditions (e.g., vegetation responses). However, models of impacts of projected changes in snow inherit the uncertainties associated with the snow projections, while the impact models have their own uncertainties. On the other hand, field experiments that manipulate snow cover according to projections have various artifacts and need to select relatively few scenarios and target a particular time horizon owing to large cost and space requirements. ***We recommend*** *improvements in the snow projection and impacts models as well as the development and deployment of new large-scale multifaceted snow manipulation experiments.*

- **Socio-economic evaluation of impacts of changing snow conditions**. It is difficult to collate information on current socio-economic impacts (Section 4.4.4) of changing snow conditions, because the impacts are usually of local importance and records are held in the relatively inaccessible grey literature of municipalities and various infrastructure authorities such as road and railway maintenance authorities. Projecting socio-economic consequences of projected changes in snow conditions is even

more uncertain as socio-economic projections must be built on projections of impacts (e.g., agriculture, forestry, reindeer herding, hydropower), each of which have uncertainties and are themselves built on projections of changing snow and climate that also have uncertainties. ***We recommend*** *an urgent focus on collating relevant information (e.g., from grey literature sources in local authorities and utility companies) and also assessing the information and making it widely available.*

- **Impacts on tourism of changing snow conditions.** There is little information about winter tourism (Section 4.4.4.6), even though this may be important for the local economy. In addition, changes in snow conditions could dramatically improve or damage such local economies depending on local circumstances. ***We recommend*** *a more detailed assessment of the consequences of changing snow conditions in the Arctic for the tourist industry.*

4.7. Conclusions: the human face of snow change

- Changes in snow conditions are likely to provide some opportunities as well as challenges. It is likely that challenges will mainly be experienced by Arctic residents while opportunities will benefit multinational industries.

4.7.1. Opportunities and challenges

Many of the impacts of a changing snow regime will have beneficial consequences for society. These include increased output of hydropower, a more even distribution of water supply and less flood damage, and reduced costs of snow clearance from roads, airstrips, etc. Current High Arctic areas with relatively little snow could become analogues of areas further south, harboring retreating wildlife and maintaining valuable scenic amenities. However, there will also be challenges associated with a changing snow regime, particularly during the transition period. Adverse impacts are already being recorded for northern commercial forests that are suffering drought because of reduced meltwater from snow in spring. Increasing frequency of extreme events in winter is also affecting forest growth, as well as reindeer husbandry, berry production, and the biodiversity and food chains of some important Arctic animals. A further challenge, and perhaps the most difficult to quantify, is the feedback from changing snow conditions to the climate system. Changes in surface albedo resulting from a different snow regime, its impacts on vegetation, and darkening by soot carried in from industrial areas at lower latitudes will lead to amplified climate warming. In contrast, the increased period of open water on lakes and longer growing seasons for vegetation will lead to greater evaporation and transpiration, drying of the landscape, and a negative feedback to climate. The balance between the two processes is highly uncertain.

4.7.2. 'Winners and losers'

The global community could suffer from the impacts of a changing snow regime in the Arctic in two fundamental ways. First, if the dominant feedbacks to the climate system from a reduced snow cover are positive, then global warming will be amplified. Second, long-term changes in snow regime and extreme events could threaten iconic species, unique snowscapes, and the perceived image of the Arctic wilderness. Within the Arctic region, urban residents are likely to benefit from many of the changes in terms of warmer winters, smaller fuel demands, more hydropower available, and less costly maintenance of transport routes. However, in the rural areas, indigenous peoples and other rural Arctic residents will face challenges to their traditional resource management and will need to diversify and adapt. These processes may in themselves be challenging due to insufficient labor pools and training.

4.7.3. Need for policy development

The residents of the Arctic can do little to mitigate global climate change and must adapt to the changes imposed on them. Policy development is therefore required outside the Arctic to strengthen global mitigation measures and within the Arctic to strengthen regional mitigation measures, such as emission

controls (from local industry, burning of garbage, marine transportation, emission control, etc.) while ensuring that the institutional structures, training, and investment are available to allow Arctic residents to adapt, pursue opportunities, and overcome challenges.

4.8. Acknowledgements

The authors are grateful to the Swedish Department of Environment, the Swedish Environmental Protection Agency (grant number 08/210), and the Swedish Research Council Formas (grant number 204-2009-45) for funding the involvement of the convening lead authors and for facilitating the involvement of Russian contributors. We also thank the other national (United States, Canada, Norway) agencies for funding the participation of their respective lead authors. Many authors contributed written text, figures, tables, and ideas, and are explicitly acknowledged on the title page, but we thank them here also. We also thank Christer Jonasson for kindly providing information on slush torrents. The AMAP SWIPA Secretariat gave valuable support throughout, and we thank the SWIPA Integration Team as well as authors of other SWIPA chapters for their pertinent discussions and encouragement. Finally, we thank the many reviewers and Dr. Barry Goodison, in particular, who provided extensive feedback.

Appendix 4.1. Existing sources for data and information on Arctic snow cover

The ability to monitor snow cover changes over the Arctic depends on the length and characteristics of available datasets and information. The most extensive surface observations are daily snow depth, but the network is still very sparse over large regions of the Arctic (Figure A1) and biased to coastal stations in the Canadian High Arctic. In addition, countries have different observing practices, and the data can be subject to various sources of inhomogeneity.

Long-term datasets are relatively sparse and have been derived from long-term meteorological stations on land and from drifting stations (Figure A2). Analyses of trends in October to May total precipitation (Figures 4.3 and 4.4 and Table 4.1) have been derived from various sectors of the Arctic:

1. Atlantic (60° – 85° N, 50° W – 30° E)
2. North European (60° – 80° N, 30° – 60° E)
3. West Siberian (60° – 80° N, 60° – 100° E)
4. East Siberian (60° – 78° N, 100° – 150° E)
5. Chukchi (60° – 73° N, 150° E – 170° W)
6. Alaskan (60° – 73° N, 170° – 140° W)
7. Canadian (60° – 85° N, 140° – 60° W)

As outlined in Table A1, there is a wide range of needs for snow-related information over a range of scales from real-time input, to weather forecast models, to climatological-scale calculations of snow loadings from annual maximum snow depth series covering 20 to 30 years. With the snow climate in the Arctic projected to become more variable in the future (ACIA, 2005), the demand for timely and accurate snow data and information is likely to increase. A comprehensive review of snow observations and data sources was provided by Key et al. (2007) and Brown and Armstrong (2008), and a summary of currently available Arctic snow data sources is provided in Table A2.

Table A1. Examples of applications requiring snow data and information.

Snow variable or derived statistics	Applications
Snow cover start and end dates	Ecological studies, climate monitoring
Snow melt onset date	Flood forecasting, transportation, ecological studies
Daily snow depth	Input to snow analysis products for numerical weather prediction (NWP) and land surface models, ground truth for satellite data, climate monitoring,

	derived statistics used in many applications
SWE (snow water equivalent) observations once per 10 to 15 days (snow course or transect)	Hydrological forecasting, water resources, validation of land surface and snow cover models
Annual maximum snow depth	Snow load calculations
Snow cover duration, depth, and density	Evaluation of frost penetration and ground thermal regime
Snow stratigraphy and layering including ice crust formation	Avalanche risk, transport, reindeer husbandry, ecology, development and validation of snow models and satellite algorithms
Albedo of snow-covered land from satellite data	Validation of climate models, study of feedback processes
Snow cover extent from satellite data	Input to NWP models, climate monitoring, ecological studies (snow cover start and end dates and duration), location of summer snow patches for reindeer herding

Table A2. Key sources for online snow-related data and information in the Arctic, August 2009.

Institution	**Snow data products online**
National Snow and Ice Data Center, Colorado	NOAA, MODIS, AMSR-E, passive microwave snow datasets; miscellaneous *in situ* datasets, including Russian snow transect data from 1966 to 2004 (Krenke, 1998). (www.nsidc.org)
PolarView, ESA	Snow cover extent mapping over Eurasian land areas of the Arctic. (www.polarview.org/services/snr.htm)
Environment Canada	Canadian historical daily snow depth and snow course data to 2003/2004. (ftp://ccrp.tor.ec.gc.ca/pub/RBrown)
Canadian Meteorological Centre	Global daily snow depth analysis at 0.3° resolution since 1998. Depth and estimated SWE (snow water equivalent) values over the Northern Hemisphere available online. (http://nsidc.org/data/nsidc-0447.html)
Canada Centre for Remote Sensing	Snow melt-off dates, snow cover fraction, and albedo of snow-covered land from AVHRR, 1982 to 2004 (Zhao and Fernandes, 2009). (http://geogratis.cgdi.gc.ca)
Rutgers University, Global Snow Lab	NOAA weekly and monthly snow cover data for the Northern Hemisphere from 1966. (http://climate.rutgers.edu/snowcover)
National Climatic Data Center, Asheville	U.S. daily snow depth data, daily snow depth observations in *Global Summary of The Day*. (http://cdo.ncdc.noaa.gov)
University of Alaska, Fairbanks	Bias-corrected precipitation and climatology for the Arctic. (www.uaf.edu/water/faculty/yang/bcp/index.htm)
European Space Agency, GlobSnow	Global weekly and monthly snow cover at 1 km resolution; daily, weekly, and monthly SWE over non-mountainous Northern Hemisphere. (http://globsnow.fmi.fi)
Russian Institute for Hydrometeorological Information, World Data Center (RIHMI-WDC)	Daily snow depth at 223 first-order meteorological stations, from the beginning of observations at each station (the earliest from 1890) until 2006. Update of snow transect data to 2005 (in progress). (http://aisori.meteo.ru/climat)
Institute of Geography, Russian Academy of Sciences	Field data on snow stratigraphy and meteorology at about 10 sites spread over northern Eurasia during several seasons, starting from 1957 until 1997. Includes ~300 snow profiles. (Planned to be available online. For now, by request from Andrey Shmakin)
Alaska Snow, Water and Climate Services	Real time and historical snow survey data. (http://ambcs.org)
Environment Yukon	Snow course data. (www.environmentyukon.gov.yk.ca/monitoringenvironment/snow_survey.php)

Indian and Northern Affairs, Water Resources Division, Canada	Historical snow course data online. (http://nwt-tno.inac-ainc.gc.ca/wrd)
WCRP CMIP3 Multi-Model Data	GCM output from CMIP3 runs (snow cover fraction, depth, SWE). (https://esg.llnl.gov:8443)
National Snow and Ice Data Center, Colorado	Daily precipitation sums at coastal and island Russian Arctic stations, 1940 – 1990 (Radionov et al., 2004b). (http://nsidc.org/data/g02164.html)
National Snow and Ice Data Center, Colorado	Meteorological data from the Russian Arctic, 1961 – 2000 (Radionov and Fetterer, 2003). (http://nsidc.org/data/g02141.html)
National Snow and Ice Data Center, Colorado	Arctic meteorology and climate Atlas (Fetterer and Radionov, 2000).

The following sections review the current state of snow cover information for the Arctic Region.

In situ observations

A comprehensive review of snow observations and data sources was provided by Key et al. (2007) and Brown and Armstrong (2008). The situation has not changed appreciably over the past decade for *in situ* snow depth and SWE observations, with the same data gaps remaining over large regions of northern Canada and Siberia. However, there has been some progress toward creating pan-Arctic snow datasets with the development of a new comprehensive dataset of surface snow observations since 1936 for Russia, the newly independent states of the former Soviet Union, and Fennoscandia (Kitaev et al., 2005; Bulygina et al., 2009).

A gridded (1° × 1°) snow depth dataset for Canada and the United States for the 1960 to 2003 period based on *in situ* daily snow depth observations was developed by Dyer and Mote (2006) and has been recently updated to 2008 (Robinson, Dept. of Geography, Rutgers The State University of New Jersey, USA, pers. comm., 2009). This dataset does not explicitly take account of topography in the interpolation process like the 1979 to 1997 ~25 km gridded snow depth and estimated SWE dataset for North America developed by Brown et al. (2003) from objective analysis. However, users should be aware that point snow depth observations are mainly made at open locations near airports and are unlikely to be representative of the surrounding terrain, especially in forested terrain where open areas accumulate less snow and melt out earlier (McKay and Gray, 1981). The daily snow-depth observing network also has large gaps over northern Canada and Siberia, and the observations are biased to coastal locations in the Canadian Arctic that can have a quite different climate compared to inland locations.

Canadian snow depth and snow course data (Brown and Brasnett, 2010) have been updated to 2003/04, but the period of data coverage is highly variable, and there are large data gaps. Site-specific snow datasets from some major research projects (e.g., SHEBA and SnowSTAR) are provided online, but there is no central repository or clearinghouse for Arctic snow data. Increasing automation of climate and weather observations has created homogeneity issues particularly for solid precipitation measurements (Yang et al., 2001). However, some progress has been made in generating pan-Arctic solid precipitation datasets corrected for bias and gauge-dependent errors (Yang et al., 2005).

Care must be taken with the interpretation of trends derived from *in situ* observations due to possible sources of bias and inhomogeneity. For example, snow course measurements in Russia prior to and after 1966 are incompatible due to procedural changes (Krenke 1998, updated 2004; State Committee on Hydrometeorology and Environment Protection of the USSR, 1985). Snow depth measurements in the former Soviet Union also have a discontinuity in or around 1950, when there was a change from taking measurements at mainly sheltered to mainly open sites. Snow depth measurements made at open sites tend to underestimate the snow accumulation compared to surrounding terrain and sheltered sites (Meshcherskaya et al., 1982).

Satellite observations

The frequency and resolution of satellite monitoring of snow cover has increased significantly since about 2000 with the MODIS, AMSR-E, and QuikSCAT satellite systems providing high-resolution information on snow cover (extent, water equivalent, and melt dates) in the visible and microwave ends of the spectrum. Microwave sensors also provide capabilities for monitoring winter thaw events and ice crust formation; for example, Derksen et al. (2010) were able to detect a regional rain-on-snow event over tundra snow cover with the AMSR-E 36.5 GHz polarization gradient due to a strong response of the horizontal polarization. The trend for increasingly finer resolution snow cover information is seen in the evolution of the NOAA hemispheric snow cover analysis, which was carried out weekly prior to 1997 and daily from 1997 onward, with the final objective being a twice-daily analysis (Helfrich et al., 2007). Over the same period, the spatial resolution of the gridded analysis product available to users has gone from ~200 km (before 1997), to 24 km (from 1997), to 4 km (from 2004).

Satellite observations in the visible and infrared spectral bands provide the most accurate estimates of snow cover extent and offer an opportunity for its mapping and monitoring at a much higher (~500 m) spatial resolution (Hall et al., 2002; Romanov et al., 2003). However, the inability to distinguish between snow-free and snow-covered land beneath the clouds causes a discontinuity both in time series and in the derived spatial distribution of snow cover and, therefore, limits applicability of these products in numerical models. Considerable progress has also been made in recent years in spatial and temporal filtering techniques to intelligently interpolate snow cover information from visible satellite data where cloud cover is obscuring the ground (e.g., Parajka and Bloeschl, 2008). Zhao and Fernandes (2009) developed and applied an intelligent interpolation scheme to produce a pan-Arctic dataset of daily snow cover fraction at a 5-km resolution from 1982 to 2004 with the AVHRR Polar Pathfinder dataset. Major advances have also been made in validating and improving high-resolution MODIS snow cover products (Hall et al., 2002; Hall and Riggs, 2007), and a monthly mean snow cover fraction product is now available from MODIS on a global 0.05° grid from September 2000 where most of the obscuring cloud cover effects have been removed (Hall et al., 2006). Similar improvements have been made in MODIS snow-albedo products to provide spatially and temporally complete data (Moody et al., 2008). Blending of satellite information from visible and infrared and the microwave spectrums (e.g., Romanov et al., 2000; Brodzik et al., 2007; Foster et al., 2008b) can also be used to overcome the cloud cover problem, but the blended map may be affected by errors in the microwave snow identification technique.

The NOAA weekly binary snow and no-snow product (Robinson et al., 1993) remains the workhorse for monitoring variability and change in northern hemisphere snow-cover extent with almost continuous weekly data from 1966 (the longest satellite-derived environmental data record). Beginning in June 1999, the weekly NOAA product was replaced by the daily Interactive Multisensor Snow and Ice Mapping System (IMS) product. The maps are digitized to a 1024 × 1024 pixel matrix at 23 km per pixel resolution and a routine has been developed to reduce these daily maps to a weekly product at the former lower resolution (Ramsay, 1998; Robinson et al., 1999). A major effort has just been completed to create a climate date record with the NOAA dataset, which involved recharting a large number of charts from the 1970s and 1980s with inconsistent mapping of patchy snow (Robinson, Dept. of Geography, Rutgers The State University of New Jersey, USA, pers. comm.). Some care must be used in documenting trends in mountainous regions during the spring/summer ablation period with these data as there is evidence that the change to higher resolution mapping of snow cover in 1999 has resulted in the mapping of smaller amounts of snow in these areas (Déry and Brown, 2007).

Satellite observations using the microwave sensor have been routinely used to map and monitor snow cover since the late 1970s (e.g., Grody and Basist, 1996; Armstrong and Brodzik, 2001; Derksen et al., 2003; Kelly et al., 2003). The ability of microwave sensors to 'see' snow through most clouds and at night is particularly attractive for monitoring snow cover over high-latitude areas. Microwave observations have some potential for providing information on snow depth, although this capability is limited to dry snowpacks only. The well-known limitations of microwave measurements consist of a relatively coarse

(currently 10 to 50 km) spatial resolution, difficulty in detecting shallow and melting snow, and an inability to distinguish between snow and cold rocks. As a result, snow cover extent in microwave-based products is frequently underestimated in flat areas, especially in spring and autumn, and is typically overestimated for mountainous areas. Passive microwave has been successful in monitoring snow melt onset and snow-off dates (Takala et al., 2009; Tedesco et al., 2009).

Estimates of SWE over the Northern Hemisphere have been derived from passive microwave sensors (SMMR and SSM/I) since 1976 (e.g., Chang et al., 1987; Armstrong and Brodzik, 2001), but the accuracy of the algorithms is generally acknowledged to be insufficient for many applications. The launch of the higher resolution (~10 km) AMSR-E passive microwave sensor in 2002 promised new advances in mapping SWE over the Arctic (Kelly et al., 2003), but progress has been slow due to a number of ongoing challenges in obtaining SWE values in the Arctic environment. Development of new boreal and tundra algorithms specific to the Arctic region (e.g., Rees et al., 2006; Derksen et al., 2009; Lemmetyinen et al., 2009; Rees et al., 2010) may overcome some of the limitations of previous global SWE algorithms. Passive microwave data have shown some skill for monitoring snow cover end dates in the Arctic (e.g., Brown et al., 2007), providing a means for monitoring variability and change in this important environmental indicator from 1978 with the combined SMMR and SSM/I data record. New high-resolution (5 km) microwave information on the frequency and duration of snow melt events has been produced from the QuikSCAT scatterometer (Wang et al., 2008). This information is available across the pan-Arctic region from 2000 and will be useful for studying phenomena sensitive to melt events. Recently Wang et al. (in prep) have combined passive microwave and QuikSCAT data to provide pan-Arctic information on melt onset anomalies over the terrestrial and marine cryosphere.

Snow cover analysis products

Numerical weather prediction requires consistent information on snow cover fraction and depth on daily to twice-daily timescales over the globe for initialization of forecast models (e.g., Brasnett, 1999). This need cannot be completely satisfied from available satellite and surface observing systems, and a number of meteorological centers are generating their own operational snow cover analysis products based on *in situ*, remotely sensed, and modeled information. The trend is for these analyses to be run at higher resolutions (1 to 5 km) and to incorporate more physically-based land surface schemes and snow models. The U.S. National Operational Hydrologic Remote Sensing Center SNOw Data Assimilation System (SNODAS, Carroll et al., 2001) is an example of this approach, which is used to provide 1-km resolution daily information on snow cover (depth, SWE, temperature, melt, sublimation) across the contiguous United States. A technique for generating continuous fields of snow cover characteristics was developed at the Institute of Water Problems of the Russian Academy of Sciences using information on snow extent and water equivalent derived from MODIS, AMSR-E, and a physical snowpack model (Kuchment et al., 2009). The European Space Agency sponsored the GlobSnow project (Kama et al., 2007), which will generate global fields of snow cover extent at 1-km resolution from assimilation of multiple satellite data sources including ERS-2, ATSR-2, Envisat AATSR, and Envisat ASAR.

Information on future changes in snow cover

Projected changes in snow cover fraction and accumulation are available for the 22 GCMs participating in the IPCC Fourth Assessment, for a range of scenarios and model ensemble runs. The scenarios have a number of limitations for Arctic applications, but do provide some idea of large-scale changes in snow cover that can be anticipated to accompany global warming. A variety of methods exist for downscaling more detailed local and regional information on changes in snow cover conditions, including high-resolution nested regional climate and snow process models, which provide detailed information on changes in snowpack properties (e.g., Bavay et al., 2009). Improved high-latitude snow cover simulations are anticipated with the next generation Earth System Models contributing to the IPCC Fifth Assessment. These will have a higher resolution, include improved land surface schemes (e.g., organic soil layers, deeper and more soil layers to resolve permafrost, improved parameterizations of snow-vegetation

interactions, and improved treatment of sub-grid scale heterogeneity), as well as dynamic vegetation and carbon cycle parameterization to provide more realistic feedbacks over northern latitudes.

Traditional knowledge

Traditional knowledge can provide important value-added content to data products and information and serves to make the data more useful and relevant to Arctic users. Development of community-based observing programs provide a mechanism for two-way knowledge transfers such as the Gwich'in First Nation IPY environmental monitoring project 'Yeendoo Nanh Nakhweenjit K'atr'ahanahtyaa – Environmental Change and Traditional Use of the Old Crow Flats in Northern Canada', which includes snow cover monitoring along with a wide range of environmental, wildlife, and cultural variables (http://classic.ipy.org/development/eoi/proposal-details.php?id=292). Similar collaborations between Sámi and scientists in Scandinavia are uniting traditional knowledge and scientific information on snow properties and monitoring (http://icr.arcticportal.org/en/ealat; Riseth et al., 2010).

Data access and archiving

Quick and easy access to Arctic snow data and information is still a challenge, although most of the satellite-related snow information can be readily searched and downloaded from NSIDC (National Snow and Ice Data Center) or PolarView. A number of operational centers also provide online access to snow products. The situation is not quite so good for *in situ* data, for which national data holdings are not always available online or are out of date. Table A2 provides a snapshot of some of the main snow data sources currently available for the Arctic region. A number of data portals have been established in recent years to facilitate access to datasets, such as Arctic Portal (http://new.arcticportal.org); Polar Data Catalogue (www.polardata.ca); Discovery, Access, and Delivery of Data for IPY (http://nsidc.org/daddi).

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5. Changing Permafrost and its Impacts

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Key Findings

- Permafrost warming, typically between 0.5 and 2 °C, generally continues in the Arctic. New data show permafrost warming has continued over the past decade in the eastern and High Arctic of Canada, in the northern Nordic regions and Svalbard, in the Russian European North and in western Siberia, although there was a reduced rate of warming during the past decade in western North America. Areas with warm, ice-rich permafrost showed lower warming rates than areas with cold permafrost or bedrock.

- An updated assessment of trends in active-layer thickness (ALT) over the past two decades shows ALT has increased at sites in Scandinavia and the Russian Arctic but surprisingly, increases in North America have only been reported from the interior of Alaska and only over the past five years.

- New projections of ground temperature throughout the pan-Arctic area suggest that by the end of the 21st century, late-Holocene permafrost in the Northern Hemisphere may be actively thawing at the southern boundary of the permafrost region and some Late Pleistocene permafrost could start to thaw at some locations.

- Regional models project that by the end of the 21st century, the upper 2 to 3 metres of permafrost will thaw over 16% to 20% of the area currently underlain by permafrost in Canada and that there will be widespread permafrost degradation over about 37% of the total area of Alaska. In Russia, increases in ground temperature of 0.6 to 1 °C by 2020 have been projected.

- Recent analyses indicate contrasting changes in hydrology in permafrost regions over past decades: landscape dryness is increasing in the boreal forest, particularly in areas of discontinuous permafrost, whereas some sub-Arctic areas are experiencing waterlogging when permafrost thaws.

- Biodiversity and ecosystem processes on land and in freshwaters are being affected by changes in hydrology. Recent studies show that the thawing of ice-rich permafrost is leading to the draining of wetlands resulting in a loss of habitat in some areas whereas in others, thawing permafrost is leading to impeded drainage and a shift in biodiversity to wetland vegetation. Thaw slumping may affect ecosystems sooner than air warming alone.

- Since the Arctic Climate Impact Assessment was published in 2005, new research has demonstrated the viability and diversity of organisms preserved in ancient permafrost.

- New measurements and estimates of soil carbon pools in permafrost indicate that there is more soil carbon in permafrost than previously thought. This carbon is likely to play a more important role in feedbacks to climate during permafrost thawing than earlier calculations suggested.

- Very high emissions of the powerful greenhouse gas nitrous oxide have recently been discovered from two locations in terrestrial permafrost regions of the Arctic. Although the importance of the process cannot be generalized across the Arctic, the strong radiative forcing potential of the gas suggests important potential contributions to climate forcing.

- High concentrations of subsea methane from the Laptev Sea have recently been recorded throughout the water column and in the atmosphere. Although more measurements are needed from other seas with different permafrost shelf conditions, the current observations reinforce concerns about a major feedback to the climate system from thawing and destabilization of subsea permafrost.

- Recent studies at some locations in the Arctic show increases in thermokarst development which are specifically important where ice-wedges degrade or palsas (peat mounds with a frozen core) decline. Permafrost degradation is also found likely to increase slope instability by increasing the amount of rockfalls and rockslides, and to increase rates of rock glacier movement. In areas with coastal permafrost, warming of permafrost may also increase coastal erosion.

- There is an increased acknowledgment and consideration of climate change and its impacts on permafrost in engineering design particularly for structures for which the consequences of failure are high. New maps based on a probabilistic approach of infrastructure susceptibility to permafrost thaw in Eurasia show that in large areas of Russia, especially in a zone along the coast, buildings and engineered structures have high vulnerability to ongoing change in climate and permafrost.

Summary

Permafrost is soil, rock, sediment or other earth material that remains at or below 0 °C for two or more consecutive years. It has an important influence on the biogeophysical environment and also on human activity. Engineering and hydrological properties of permafrost as well as landscape hydrology, geomorphology and biological processes will change if permafrost warms and thaws. Thawing permafrost will affect populations locally and globally.

Current and projected changes in permafrost conditions are likely to result from the interaction of several factors and processes, particularly increasing air temperatures that are generally leading to permafrost degradation. However, local factors are also important, including snow cover, vegetation, organic layer thickness, thermal properties of the earth material, soil moisture/ice content and drainage conditions. Changes in these local factors through natural processes or environmental disturbance associated with human activity can lead to changes in permafrost conditions.

Observations from permafrost monitoring sites throughout the Arctic, including North America, Russia and Scandinavia generally indicate that permafrost temperatures have increased, typically by 0.5 to 2 °C, although the rate and magnitude of the increase varies regionally. Areas with warm ice-rich permafrost showed lower warming rates than areas with cold permafrost or bedrock. Monitoring of active-layer thickness (ALT) indicates decadal trends that vary by region. A progressive increase in ALT has been observed in some Nordic sites, the Russian European North, East Siberia, and Chukotka with disappearance of permafrost from the upper several metres over the past 20- to 30-year period at several low Arctic Nordic, Canadian and European Russian sites. North American sites show a progressive increase of ALT during the last five years only at a limited number of sites in the Alaskan Interior, and ALT on the Alaskan North Slope and in the northwestern Canadian Arctic has been relatively stable since 1995.

Permafrost models, driven by projections of Arctic warming by general circulation models (GCMs), project a 16% to 20% reduction in the areal extent of near-surface (depth 2 to 3 m) permafrost in Canada by the end of the 21st century. Between 9% and 22% of remaining permafrost in Canada would contain taliks (unfrozen ground within the permafrost). In Alaska, mean annual ground temperature (MAGT) at 2

m depth could be above 0 °C everywhere southward of 66° N except for small patches at high altitude in the Alaska Range and Wrangell Mountains. An area of approximately 850 000 km^2 (about 57% of the total area of Alaska) is projected to experience widespread permafrost degradation and could contain areas where permafrost disappears completely. In Russia, permafrost changes of a +0.6 to +1 °C increase in ground temperature have been projected by 2020.

When permafrost starts to thaw, waterlogging occurs in some predominantly flat areas, whereas ponds dry in others. These opposing outcomes of thawing permafrost have profound impacts on infrastructure, ecology, and greenhouse gas emissions. In northern mountain areas thawing permafrost can also lead to rock slope instability. Also, coastal erosion will increase as permafrost thaws because much of the Arctic coastline is composed of unconsolidated materials, many of which are ice-rich and sensitive to increased air temperature and changes in the length of seasonal sea-ice cover.

Since the publication of the Arctic Climate Impact Assessment (ACIA, 2005) and the Fourth Assessment of the Intergovernmental Panel on Climate Change (e.g., Solomon et al., 2007), knowledge about the preservation and activity of life in permafrost has increased. Furthermore, recent work has shown that carbon pools in permafrost soils are much larger than previously recognized: around 1400 to1850 gigatonnes (Gt) of carbon are located in terrestrial permafrost regions. Although model projections suggest that tundra is likely to remain a weak sink of atmospheric carbon dioxide (CO_2), there are great uncertainties and the emissions of methane (CH_4) and nitrous oxide (N_2O) (much stronger greenhouse gases than CO_2) from permafrost areas have the potential to substantially increase radiative forcing. In addition, Arctic coastal seas underlain by subsea permafrost host an extremely large carbon pool: the Arctic continental shelf could contain around 1300 Gt of carbon, of which 800 Gt is CH_4, some of which could be available for sudden release under the appropriate conditions. A release of only 1% of this reservoir would more than triple the atmospheric mixing ratio of CH_4, potentially triggering abrupt climate change.

Increasing ground temperature and thawing permafrost can cause ground subsidence in areas of ice-rich permafrost. Infrastructure design takes into account that there may be some degree of permafrost thaw in response to changes in the ground thermal regime associated with construction activities and infrastructure operation. Climate change is an additional factor that needs to be accounted for, and in some areas this is already being done for design of major structures.

Climate feedbacks from thawing permafrost in the Arctic can help to stimulate a global response to climate change. However, mitigation by local communities will be relatively insignificant and they must therefore adapt to the changing conditions. Adaptation through engineering is already taking place and appropriate technology is available, but costly. To formulate and implement other adaptation strategies a range of tools must be developed. These include hazard index calculations, high spatial resolution projections of permafrost change and probabilistic active-layer change maps.

Thawing permafrost offers few opportunities but many challenges. Large multinational industries and developers will face the need for greater economic investment to stabilize infrastructure over longer periods, whereas individual residents will face disruption to communication routes and even resettlement in some cases. However, when permafrost completely disappears at its southern margins, communities will benefit as they will no longer require special infrastructure design features related to permafrost.

5.1. Introduction

Permafrost is soil, rock, sediment or other earth material that remains at or below 0 °C for two or more consecutive years (Brown and Péwé, 1973). It underlies most of the surfaces of the terrestrial Arctic (French, 2007) while coastal and subsea permafrost exist around Arctic coastlines.

Progressing from the soil surface downward, there is an active layer which freezes and thaws seasonally, a transient layer that can remain frozen in some summers (Shur et al., 2005), and then permafrost (Figure 5.1). Most subsea permafrost is not subject to the seasonal variation seen on land, since its upper temperature is controlled by bottom seawater temperatures. As long as sediment temperatures are cryotic (i.e., less than 0 °C), the sediment is classified as permafrost, whether or not it contains ice. Unfrozen zones (taliks) can occur within permafrost, for example under large water bodies. Terrestrial permafrost thickness ranges from a few tens of centimeters at the southern limit of the permafrost zone to about 1500 m in the north of the Arctic region. Active-layer thickness is influenced by climate and local factors and can vary from less than 0.5 m in vegetated, organic terrain to more than 10 m in areas of exposed bedrock.

The proportion of the landscape underlain by permafrost generally becomes greater with increasing latitude from the southern limits of the permafrost zone to the High Arctic. This transition is reflected in isolated islands and sporadic permafrost at the southern margins, more extensive but discontinuous permafrost further north, and more or less continuous permafrost found everywhere except beneath large bodies of water or newly aggraded land in the High Arctic (Brown et al., 1998; Figure 5.2).

Permafrost has an important influence on the biogeophysical environment and also on human activity, largely because it often contains ice. The engineering and hydrological properties of earth material, as well as its influence on biological processes will therefore change if permafrost warms and thaws. The strength and stability of frozen ground is strongly related to temperature and can decrease as ground ice melts. This has major implications for infrastructure performance (Instanes et al., 2005), especially if the ground is ice-rich. Permafrost also influences surface and subsurface hydrology as frozen ground restricts the mobility of groundwater. Changes in the permafrost can lead both to drying and waterlogging of the landscape which has important implications for water sources, infrastructure, ecology, and biogeochemical cycling (e.g., land/air exchange of greenhouse gases). Permafrost moderates properties of the active layer that affect greenhouse gas fluxes and also contains vast reserves of historically sequestered carbon (Tarnocai et al., 2009) that, if released, could act as a positive feedback to climate warming.

Climate, in particular air temperature, is one of the main factors influencing the areal extent and thickness of permafrost as well as its thermal condition. The projected future increase in air temperature is therefore expected to result in changing permafrost conditions. However, local factors are also important in influencing the energy balance at the ground surface and the response of permafrost to changes in climate (Williams and Smith, 1989; Yershov, 1998; French, 2007). These factors include snow cover, vegetation, soil organic layer thickness, thermal properties of the earth materials, soil moisture/ice content and drainage conditions. Changes in these local factors through natural processes or environmental disturbance associated with human activity can thus lead to changes in permafrost conditions. Forest fires for example, can cause removal of vegetation cover and the insulating organic layer and result in warming and thawing of permafrost. Shifting shorelines resulting from river erosion and sediment deposition can also result in permafrost degradation as areas become inundated. However, permafrost formation can also occur as formerly flooded areas are exposed to lower air temperatures. Permafrost formation can also occur in newly exposed land such as the bottom of drained lakes (e.g., Mackay and Burn, 2002), in deglaciated areas in front of the Greenland Ice Sheet and local glaciers, and in areas of coastal uplift or sedimentation such as deltas (e.g., Dyke et al., 1997; Taylor, 1991). At a pan-Arctic scale, permafrost formation at present is relatively rare, but is locally important.

Human activity associated with the construction and operation of infrastructure can also lead to changes in the ground surface and alterations in the ground thermal regime. Land clearance and removal of organic matter during construction activities, result in a reduction in the buffer layer and warming of the ground, increases in thaw depth, and possible degradation of permafrost. The placement of heated structures on (e.g., buildings) or in (e.g., pipelines) permafrost can also result in thawing.

Not all areas underlain by permafrost are equally vulnerable to short-term thaw. Southern areas underlain by ice-rich permafrost and areas of sporadic and discontinuous permafrost are likely to be most impacted

by increased air temperature. In contrast, the cold thick permafrost of Siberia and the North American High Arctic might experience modest surface warming but the massive frozen materials at depth are likely to be stable for millennia. Such variation in the sensitivity of permafrost to climate warming differentially affects the vulnerability of ecological processes and human activities to change. Adaptation of biological systems to changes in permafrost is an ongoing process and is part of a two-way interaction between vegetation and permafrost. However, human adaptation to changes in permafrost is a relatively new dimension that involves multiple aspects of land use and infrastructure development that incur cultural and economic costs, rather than new opportunities.

This chapter provides a synthesis of current knowledge of existing and projected changes in Arctic permafrost conditions and their likely impacts on society. This adds to the recent benchmark assessments of the Arctic Climate Impact Assessment (ACIA, 2005) and the Fourth Assessment of the Intergovernmental Panel on Climate Change (e.g., Solomon et al., 2007) by focusing on more recent information and new topics. The chapter reports recent changes in permafrost conditions and synthesizes new projections of future change and its implications for natural systems and society. This information is used to highlight how adaptations to changing permafrost conditions might occur in the natural world and how society might develop socio-economic adaptation strategies. The chapter concludes by listing important gaps in understanding and recommends priority actions. Many types of knowledge sources are drawn on (e.g., field observations and monitoring, experimental and modeling activities), with the focus on new data acquired during the International Polar Year projects and output from the Ninth International Conference on Permafrost (NICOP) in 2008 and the Third European Conference on Permafrost (EUCOP III) in 2010.

5.2. Past, current and future states of permafrost and causes of change

- Permafrost warming, typically between 0.5 and 2 °C, generally continues in the Arctic. New data show permafrost warming has continued over the past decade in the eastern and high Arctic of Canada, in the northern Nordic regions and Svalbard, in the Russian European North and in western Siberia, although there was a reduced rate of warming during the past decade in western North America. Areas with warm, ice-rich permafrost showed lower warming rates than areas with cold permafrost or bedrock.

- An updated assessment of trends in ALT over the past two decades shows ALT has increased at sites in Scandinavia and the Russian Arctic, but surprisingly, increases in North America have only been reported from the interior of Alaska and only over the past five years.

- New projections of ground temperature throughout the pan-Arctic area suggest that by the end of the 21st century, late-Holocene permafrost in the Northern Hemisphere may be actively thawing at the southern boundary of the permafrost region and some Late Pleistocene permafrost could start to thaw at some locations.

- Regional models project that by the end of the 21st century the upper 2 to 3 m of permafrost will thaw over 16% to 20% of the area currently underlain by permafrost in Canada and that there will be widespread permafrost degradation over about 57% of the total area of Alaska during this period. In Russia, increases in ground temperature of 0.6 to 1 °C by 2020 have been projected.

5.2.1. Permafrost in the past

Knowledge about past permafrost distribution and dynamics leads to a better understanding of the significance and scale of recent changes in the thermal state and distribution of permafrost, and this knowledge-base has increased recently. It also facilitates the projection of possible rates and pathways of future permafrost degradation. There are two principal reasons for this. First, the main present-day features of permafrost distribution, both vertically and laterally, were formed during the past 100 000 years. Second, with persistent future climate warming, changes in current permafrost will reflect its

history: the first permafrost to begin to thaw will be the youngest Little Ice Age permafrost, followed by mid- and late-Holocene permafrost, and lastly, Late Pleistocene permafrost.

5.2.1.1. Past climate and permafrost development

Permafrost is a product of a cold climate. Thus, in the past, whenever and wherever the Earth's climate became cold enough, permafrost has developed (Yershov, 1998). Progressive cooling of the Earth's climate, which started about 40 million years ago, caused expansion of the surface area occupied by ice masses and permafrost. Since then, the Earth's climate has been generally conducive to the existence of permafrost at high latitudes and high elevations in both the Northern Hemisphere and Southern Hemisphere.

Huge changes in climate and other environmental characteristics including northern hemisphere permafrost distribution occurred across the northern high latitudes during the Pleistocene (roughly the past 1.5 million years; see also Figure 5.3). These changes were driven by the Milankovitch cycles in solar radiation received by the Earth's surface (Milankovitch, 1930; Imbrie and Imbrie, 1979; Melnikov and Smulsky, 2009). Climate variations on the glacial-interglacial time-scale were also responsible for dramatic changes in global sea level. During the last glacial maximum (around 20 ky BP), sea level was 120 to 140 m lower than at present. Thus, practically all present-day Siberian Arctic shelves and a significant proportion of the North American Arctic shelves (e.g., the Beaufort) were dry land that was exposed to extremely cold climatic conditions (Hubberten and Romanovskii, 2001; Overduin et al., 2007; Velichko and Faustova, 2009; Velichko and Nechaev, 2009; see also Figure 5.3). During this period, terrestrial permafrost up to 500 m thick was formed on the inner shelf areas, but was thinner towards the shelf edge due to the shorter exposure time of land to the cold air.

Climate was also the major factor affecting permafrost during the transition periods from glacial to interglacial conditions and during the interglacial periods. However, within low-lying Arctic coastal areas, sea-level rise was also a major driving force of permafrost degradation. During these warm periods, other environmental changes, such as changes in hydrology and vegetation, started to play increasingly important roles in the preservation or degradation of permafrost until eventually, changes in permafrost on the millennia and century time-scales were driven as much by changes in vegetation and hydrology as by changes in climate (Velichko and Nechaev, 2005; Jorgenson et al., 2010).

5.2.1.2. Magnitude and rate of changes

During the last glacial maximum, permafrost underlay more land area than today (see Figure 5.3). The climatic transition from the last glacial period to the current interglacial period was associated with a rapid thaw of permafrost both from the top and bottom at the southernmost limits of its Late Pleistocene maximum distribution, although some permafrost formation also occurred (e.g., in areas of coastal uplift). With climate warming in progress, rapid permafrost degradation in the Northern Hemisphere became more extensive. The rate of permafrost thawing, which varied from a few millimeters to several tens of centimetres per year, depended to a significant extent on permafrost ice content and the texture of thawing sediments. By the time of the Holocene optimum (5 to 9 ky BP; see Figure 5.3), permafrost had completely disappeared from most of the territory of deglaciated Europe, from northern Kazakhstan, and from a significant proportion of western Siberia in northern Eurasia (Velichko and Nechaev, 2009; Velichko and Faustova, 2005). In areas where the upper several hundred metres of permafrost was ice-rich, such as in the Pechora River basin and in the northern and central parts of western Siberia, permafrost did not disappear completely and is still present at great depth (200 m and deeper) (Balobaev et al., 1983; Baulin, 1985; Melnikov and Grechishchev, 2002) (Figure 5.4).

During the Holocene optimum, much of the continuous terrestrial permafrost zone in the Arctic, was generally stable with no widespread thaw. However, where near-surface permafrost was ice-rich, many thermokarst lakes developed causing localized thawing under those lakes that were sufficiently deep

(MacDonald et al., 2006; Walter et al., 2007a). The rate of thawing both in the vertical and lateral direction was several centimetres to several tens of centimetres per year. Several thousand years after the Holocene optimum, permafrost started to thaw from the bottom up in this continuous permafrost zone (Osterkamp and Gosink, 1991), but the rate of thaw was only a few centimetres per year.

Holocene climate has been generally, much more stable and warmer than during the Late Pleistocene. However, there were several relatively cold intervals that lasted several centuries in the Middle and Late Holocene (6 to 2 ky BP: Velichko and Nechaev, 2005). During these periods, new, fairly shallow, short-lived permafrost appeared and disappeared several times in some landscape types found within the sporadic and discontinuous permafrost zones near the southern boundary of the present-day permafrost (Figures 5.2 and 5.4). The last and probably the coldest of such intervals was the Little Ice Age that dominated most of the Northern Hemisphere climate between around 1600 and 1850. During this period, shallow permafrost (15 to 25 m, e.g., Romanovsky et al., 1992) was established within the sediments that had been predominantly unfrozen during most of the Holocene. Present-day warming initiated the Little Ice Age permafrost thawing that is ongoing today (see Section 5.2.2.2). In the High Arctic, the reconstructed ground surface temperature record for a small island site shows that there was a muted response of permafrost to the Little Ice Age due to the maritime climate, compared to sites influenced by a more continental regime (Taylor et al., 2006). These reconstructions also show that cyclical fluctuations in permafrost temperature, related to decadal cycles in atmospheric circulation patterns are superimposed on the longer-term warming of permafrost since the Little Ice Age.

5.2.2. State of present-day terrestrial and subsea permafrost and controlling factors

5.2.2.1. Distribution of permafrost

Permafrost areas presently occupy 23% to 25% of the terrestrial parts of the Northern Hemisphere, most of which is in the Arctic (Brown et al., 1998). Continuous permafrost dominates the northernmost land areas while its distribution becomes discontinuous in the sub-Arctic landscape (Figure 5.2). However, continuous permafrost extends into the boreal forest (taiga) regions with a continental climate in some areas (e.g., Burn and Kokelj, 2009; and compare permafrost distribution in Figure 5.2 with forest distribution in figure 1 of Callaghan et al., 2002). Permafrost is divided into four zones based on the percentage of land that it underlies: continuous permafrost (90% to 100%), discontinuous permafrost (50% to 90%), sporadic permafrost (10% to 50%), and isolated islands of permafrost (0% to 10%) (Brown et al., 1998).

Permafrost thickness reaches a maximum of 1500 m in Siberia (French, 2007) in areas which were not covered by glaciers for substantial periods, whereas it is generally less than 400 m thick in areas covered by glaciers during the last glaciation, such as in North America (French, 2007). Very detailed permafrost mapping exists for Russia and Alaska, while Canadian mapping is less detailed, and European and Greenland permafrost is only mapped in low detail. The circum-Arctic permafrost map (see Figure 5.2) also includes some information on the distribution of key permafrost landforms (see also Section 5.3.2.1), but there is no complete systematic mapping of permafrost landforms at this scale.

The occurrence of subsea permafrost has been demonstrated by several geophysical studies involving drilling on the Siberian, Chukchi and Beaufort shelves but data are mostly unpublished. Subsea permafrost occurs in the continental shelves of the Arctic Ocean adjacent to Russia (Figure 5.5), Alaska, Canada, Greenland, and Svalbard (Baranov, 1960; Vigdorchik, 1980; Osterkamp, 2001) where it is likely to exist beneath depths of water of up to 100 m (Mackay, 1972; Osterkamp and Harrison, 1982; Osterkamp and Fei, 1993; Hinz et al., 1998). Due to changes in sea level and to shoreline erosion, the present configuration of subsea permafrost in the Beaufort Sea resembles a wedge or tabular sheet extending from the Arctic Ocean coast down to 100 m water depth, a distance of up to 100 km (Osterkamp and Fei, 1993). On the huge Siberian shelf, the 100 m water depth may be over 700 km offshore (Romanovskii et al., 1998).

5.2.2.2. Recent changes in permafrost temperature, and active-layer thickness and extent

5.2.2.2.1. Recent changes in permafrost temperature

A recent coordinated field campaign conducted during the International Polar Year (2007/2008) collected permafrost temperature data throughout the Arctic (Romanovsky et al., 2010b) and enhanced the existing permafrost monitoring network by establishing more than 300 new permafrost boreholes. This recent snapshot (Christiansen et al., 2010; Romanovsky et al., 2010a; Smith et al., 2010; Figure 5.6) shows a general decrease in permafrost temperature with increasing latitude although this relationship varies regionally. For example, the influence of warm ocean currents on the climate of northern Scandinavia, Svalbard and northwestern Russia (McBean et al., 2005) results in higher permafrost temperatures in these regions compared to other High Arctic regions at a similar latitude. In the discontinuous permafrost zone, permafrost temperatures fall within a narrow range, with mean annual ground temperatures at the level of zero annual amplitude being generally warmer than -2 °C. However, there is still much spatial variability within this zone due to the variability of local factors such as snow cover, vegetation and the presence of an insulating (organic) layer (Smith et al., 2010; Romanovsky et al., 2010a). A greater range in permafrost temperature is found within the continuous permafrost zone where mean annual ground temperature can be as low as -15 °C in the High Arctic, although permafrost at temperatures below -10 °C is only presently found at monitoring sites in the Canadian Archipelago and northern Russia.

Over the past two to three decades, there has been a general warming of permafrost across the circum-Arctic permafrost region (Lemke et al., 2007; Romanovsky et al., 2007; Brown and Romanovsky, 2008). Observations from permafrost monitoring sites throughout the Northern Hemisphere, including North America (e.g., Smith et al., 2005a, 2010; Osterkamp, 2008), Russia (e.g., Malkova, 2008; Oberman, 2008; Romanovsky et al., 2008a, 2010a) and Scandinavia (Harris and Isaksen, 2008, Christiansen et al., 2010) generally indicate that the magnitude of the increase in permafrost temperature varies regionally. The magnitude has typically been between 0.5 and 2 °C at the depth of zero annual amplitude since the late 1970s (Figure 5.7). Warming rates have been much lower for warm ice-rich permafrost at temperatures close to 0 °C than for colder permafrost (< -2 °C) or bedrock (Romanovsky et al., 2010b). Overall, the range in permafrost temperature is about 1 °C smaller than 30 years ago (Romanovsky et al., 2010b).

Variability in the rate of temperature change is exemplified in Alaska and the Mackenzie valley region of the Northwest Territories Canada, a region which has experienced an increase in air temperature of about 1.5 °C over the past century. Permafrost temperatures at depths of 10 to 20 m in this region have increased by a few tenths of a degree per decade in the south and central area and about 1 °C per decade in the north (e.g., the Arctic Coastal Plain, Mackenzie Delta region) since the 1980s (Kanigan et al., 2008; Osterkamp, 2008; Romanovsky et al., 2008b; Smith, et al., 2005a, 2010; Burn and Kokelj, 2009). Ground temperatures measured recently at Herschel Island, Yukon, and associated modeling suggest that permafrost temperature at 20 m depth may have increased by 1.9 °C over the past century (Burn and Zhang, 2009). However, there has been a decrease in the rate of permafrost temperature increase in western North America since 1998 (Smith et al., 2010), a year with the warmest air temperatures on record for this region. In general, air temperatures have been lower since 1998 and the ground temperatures have shown a similar pattern. The smaller increases in permafrost temperature observed in the more southerly areas where permafrost is thinner and at temperatures close to 0 °C, are due to latent heat effects, associated with phase changes (thawing and freezing), that dominate the ground thermal regime (especially where permafrost is ice-rich) (Smith et al. 2005a, 2010; Romanovsky et al., 2010b).

In the colder permafrost (about -15 °C) of the High Arctic tundra, ground temperatures are more responsive to changes in air temperature due to the lack of a buffer layer (little vegetation, thin snow cover) and the lack of phase changes. At Alert, Nunavut, permafrost at a depth of 15 m warmed by about 0.1 °C per year between 1994 and 2002 (Smith et al., 2003, 2005a) in response to warmer air temperatures and more recent observations indicate that this trend is continuing (Smith et al., 2010). These observations

are consistent with those of other Arctic areas such as Scandinavia and Svalbard where recent increases in shallow ground temperatures have also been observed (Isaksen et al., 2007a,b; Harris and Isaksen, 2008; Johansson et al., 2008; Christiansen et al., 2010).

An increase in climatic variability (both extremely cold and extremely warm years) occurred in European Russia during the past decade. This led to a considerable increase in permafrost temperatures. In the Russian European North and in northwestern Siberia, temperatures at the depth of zero annual amplitude generally increased between 1974 and 2007. Increases of up to 2 °C occurred at colder permafrost sites and up to 1 °C at warmer permafrost sites (Brown and Romanovsky, 2008; Oberman, 2008; Romanovsky et al., 2008b, 2010a).

The very limited data available show that permafrost is actively degrading within submerged Arctic shelves (Zhigarev, 1997; Rachold et al., 2007) under both the thermal and chemical impact of overlying seawater. However, the rate and exact location of this degradation are mostly not known.

5.2.2.2.2. Recent changes in active-layer thickness

Recent results from high-latitude Arctic sites within the Circumpolar Active Layer Monitoring (CALM) program (see Appendix 5.1), indicate substantial interannual fluctuations in ALT, primarily in response to variations in summer air temperature (e.g., Smith et al., 2009a; Popova and Shmakin, 2009; Figure 5.8). Decadal trends in ALT vary by region. A progressive increase in ALT has been observed in some Nordic sites (e.g., Akerman and Johansson, 2008), northeastern Greenland (Christiansen et al., 2008), the Russian European North (e.g., Oberman and Mazhitova, 2001; Mazhitova et al., 2008), eastern Siberia (e.g., Fyodorov-Davydov et al., 2008), and Chukotka (e.g., Zamolodchikov et al., 2008). Disappearance of permafrost over the past few decades has been reported at several lower latitude Arctic sites, including Nordic (e.g., Akerman and Johansson, 2008), Canadian (e.g., Smith et al., 2008a), and European Russian (e.g., Mazhitova et al., 2008; Oberman, 2008) sites, that are mainly peatlands in the sporadic and discontinuous zones. In eastern Siberia, the observed changes in ALT have resulted in degradation of the ice- and organic-rich transient layer on top of permafrost (Shur et al., 2005).

Trends in ALT are different for North American sites. A progressive increase in ALT is evident only at sites in the Alaskan interior, where in 2007 the maximum ALT for the 18-year observation period occurred (Viereck et al., 2008). Active-layer thickness at the Alaskan North Slope and northwestern Canadian Arctic sites is relatively stable, without pronounced trends during the 1995 to 2008 period (e.g., Streletskiy et al., 2008; Smith et al., 2009a).

Active-layer thickness varies between years and extremely warm years can enhance increase in ALT. In the maritime High Arctic Nordic CALM site in Svalbard, interannual variation was as high as 30 cm during an eight-year period (Christiansen and Humlum, 2008). Canadian monitoring sites showed responses in ALT, consistent with those of other cryospheric components, to the extreme warming in 1998 that resulted in a longer thaw season (Atkinson et al., 2006). The greatest increases in thaw occurred at the more northerly sites, especially those lacking a vegetation cover or an organic surface soil layer (Smith et al., 2009a).

5.2.2.2.3. Recent changes in permafrost extent

A new study, published since the Arctic Climate Impact Assessment in 2005 (ACIA, 2005), has addressed regional permafrost changes in the Russian European North (Oberman and Liygin, 2009). This covered a wide area and included a variety of bioclimatic, topographic and permafrost conditions as well as all three permafrost zones, while providing unique information on recent dynamics of the frozen ground during the 1970 to 2005 period.

Observations during the 1970 to 2005 period at some sites in this region indicate significant (by several metres) thawing of the permafrost in the uppermost near-surface layer. In addition to these changes in permafrost thickness, the area underlain by permafrost has also decreased. The permafrost boundary shifted northward by 30 to 40 km in the Pechora lowlands and up to 80 km on the foothills in the eastern part of the region (Figure 5.9). Large areas in the discontinuous and southernmost part of the continuous permafrost zones have been affected by thawing. This has led to deepening of existing taliks and to the formation of new ones. As a result, the boundary of continuous permafrost has moved northward by 15 to 20 km in the lowlands and by 30 to 50 km in the foothills.

Similar observations have been made in eastern Canada (Quebec) where the southern boundary of the permafrost has shifted northward by 130 km during the past 50 years (Thibault and Payette, 2009). Permafrost has degraded since 1957 at sites in the sub-Arctic peatlands of northern Quebec, with the rate of loss increasing to 5.3% per year since 1993 (Payette et al., 2004). In one location, the area occupied by palsas decreased by 23% between 1957 and 2001 (Vallée and Payette, 2007). In western Canada, permafrost has degraded over the last 100 to 150 years at five southern sites, with the degradation ranging between 30% and 65%, while as much as 50% of permafrost in peat plateaus thawed over 50 years at four sites in the northern part of the discontinuous permafrost zone of the Mackenzie Valley (Beilman and Robinson, 2003). In Tanana Flats, central Alaska, areas with thermokarst fens and bogs increased from 39% to 47% of the area over 46 years (Jorgenson et al., 2001).

5.2.2.3. Factors controlling the current state and dynamics of permafrost

5.2.2.3.1. Terrestrial permafrost

Changes in permafrost temperature and active-layer thickness are generally consistent with trends in air temperature. While past air temperature has been important for the formation of permafrost, current air temperature is important for the continuing existence of permafrost, since air temperatures below freezing are required to keep the ground from thawing. Variations in snow cover are also an important driver of permafrost temperature: warming of permafrost in the late 1980s and early 1990s in Alaska, for example, may be mainly due to a series of thicker snow covers (Osterkamp, 2008). In the High Arctic, changes in snow cover were found to counteract changes in air temperature. As a result, permafrost temperature may rise during periods of lower air temperature if more snow occurs (Taylor et al., 2006).

The presence of a surface buffering layer (including a peat layer) as well as ice-rich material can be important factors that lead to persistence of permafrost through warming events. For example, ground ice with an age of 740 ky BP found in the central Yukon Territory (northwestern Canada) may have persisted within the discontinuous zone through several interglacial cycles and exposure to warmer climates (Froese et al., 2008). Permafrost at temperatures close to 0 °C is also observed in the peatlands of the southern part of the permafrost zone that has persisted through the warming following the Little Ice Age, due to the insulation provided by the peat (Smith et al., 2008a; Kokfelt et al., 2010). The importance of organic layers influencing the response of permafrost to climate change has also been explored in modeling studies (Woo et al., 2006b, 2007; Yi et al., 2007; Lawrence et al., 2008). Model results suggest that the increase in thaw will be lower in response to climate warming in areas with an insulating organic layer. Thus, peat could preserve permafrost against the projected warming in many northern areas. However, higher resilience of permafrost protected by a thick organic layer can be threatened by various natural and human-induced surface disturbances, such as forest, tundra, or peatland fires and agricultural or engineering activities (e.g., Viereck et al., 2008; Smith et al., 2008a). Surface geomorphological processes, such as erosion, thermal erosion, various slope processes, and thermokarst, may also destroy this protective organic layer and dramatically accelerate permafrost degradation in the areas affected (Kokelj et al., 2009; Jorgenson et al., 2010).

5.2.2.3.2. Subsea permafrost

Most Arctic subsea permafrost is inundated relict terrestrial permafrost. There are some exceptions, for example the refreezing of freshwater river discharge onto the cryotic seabed, or the formation of permafrost in areas such as deltas, where sediment accumulation raises the seabed close to the water surface, permitting bottom-fast ice formation in winter. Since this depends on local conditions, these exceptions are limited in spatial extent.

Once the previously coast is inundated, ground temperature and salinity at the seabed are controlled by sea ice dynamics and the temperature regime of the sea bottom water. In most locations, the sea-ice dynamics are assumed to be relatively stable at seawater temperatures just below 0 °C.

5.2.3. Projections of future permafrost states

Global circulation models project a 1.4 to 5.8 °C increase in global air temperature during the 21st century (ACIA, 2005) with an increase of 3 to 6 °C projected for the Arctic by 2080 (see Chapter 3, Section 3.2). This warming would induce permafrost degradation (i.e., thawing of the frozen ground, thickening of the active layer, and a reduction in spatial extent).

Permafrost projections differ in geographical extent and spatial resolution, and depend largely on the resolution of the forcing climatic and environmental parameters (soil, vegetation etc.), and on the permafrost model used. Several types of permafrost model of different complexity, computational design, and data requirements have been developed and used in various applications. A survey of permafrost modeling approaches was reported by Riseborough et al. (2008). See also Box 5.1.

Box 5.1. Permafrost modeling approaches

Permafrost models link climate, particularly air temperature and precipitation, with the state of the frozen ground, which is typically characterized by ground temperature and the depth of maximum seasonal thawing (active-layer thickness). Soil and vegetation properties largely govern this relationship, and most permafrost models include them explicitly.

There are two conceptually different modeling approaches: one based on 'equilibrium models' and the other based on 'transient models'. Equilibrium models are relatively simple and assume that the ground thermal regime is in equilibrium with the atmospheric climate thus neglecting the time lag between climatic forcing and its impact on the frozen ground. Such models have relatively low data requirements, using mean monthly temperature and precipitation data as climatic forcing together with a few edaphic parameters characterizing soil thermal properties, snow and vegetation. Owing to their computational design and inherent limitations, such models may only be used to describe active-layer thickness and temperature in the uppermost near-surface permafrost layer. Despite these limitations, equilibrium models have been employed in many studies to construct permafrost projections at a variety of geographical scales, from regional to hemispheric (Sazonova et al., 2004; Anisimov and Reneva, 2006; Sushama et al., 2006; Saito et al., 2007; Duchesne et al., 2008; Marchenko et al., 2008; Romanovsky et al., 2008a). Results obtained with this type of model are referred to as 'equilibrium permafrost projections' later in this chapter.

More advanced, transient models simulate the dynamics of the ground temperature regime and seasonal thawing/freezing, so that transient responses of permafrost to changing climatic and environmental conditions may be projected. Computationally, they are much more expensive than equilibrium models, and in most cases require additional climate forcing data, such as solar radiation, wind, cloudiness, and air humidity at a daily or monthly resolution. The most sophisticated models of this type include dynamically coupled blocks that calculate changes in soil moisture and thermal properties at every time step rather than using prescribed constant soil-specific values for these parameters, as is the case in equilibrium model calculations. Although transient models have high predictive capacity, their application is limited,

particularly over large regional and hemispheric scales, because many of the necessary forcing data are often unavailable (Zhang et al., 2008a,b).

An inherent limitation of conventional equilibrium and transient models is their deterministic nature, which contrasts with actual permafrost parameters that are intrinsically stochastic and largely governed locally by small-scale variability in soil, vegetation, snow parameters, and topography.

Many potential applications of permafrost models, such cold-region engineering, require probabilistic metrics. A new type of stochastic permafrost model has thus been developed recently, this is known as a 'probabilistic permafrost model'. Unlike conventional models, they calculate statistical ensembles representing potential states of permafrost under prescribed conditions (Anisimov et al., 2002; Anisimov, 2009). This new methodology is fully harmonized with the ensemble approach that is used to construct probabilistic climatic projections based on results derived from several GCMs. However, it also provides important information that directly addresses the practical needs of stakeholders and may be used in various applications such as the risk assessment of potential infrastructure damage, evaluations of other threshold-driven processes, and impacts associated with thawing permafrost.

The approaches described above are based on stand-alone permafrost models and require independent climatic data. Another approach is to incorporate permafrost directly into GCMs by improving parameterization of the related ground processes (Stendel and Christensen, 2002; Lawrence et al., 2008; Wania et al., 2009): these models are known as 'coupled models'. However, these generally only consider the upper few metres (3 to 6 m) of the ground and do not adequately consider time lag effects (Burn and Nelson, 2006). This can lead to significant overestimates of permafrost thaw (Burn and Nelson, 2006; Delisle, 2007) as deeper ground should be considered for a more realistic projection (Nicolsky et al., 2007; Lawrence et al., 2008).

A special group of models is that used for mountain permafrost projections. In complex mountainous terrain, vertical and horizontal heat fluxes both play an equally important role in the thermal regime and modeling is undertaken in three-dimensional space.

Over the past decade, several studies have projected the impacts of climate warming on permafrost in the 21st century. Most ACIA (2005) projections are equilibrium projections as they assumed that the ground thermal regime is in equilibrium with the current atmospheric climate (Box 5.1). A major recent advance has been to simulate ground temperature dynamics such that the transient changes of permafrost can be projected (Box 5.1). The following sections report Arctic-wide and regional projections.

5.2.3.1. Arctic-wide permafrost projections

The GIPL2/MPI (Geophysical Institute Permafrost Lab, University of Alaska Fairbanks) parallel transient model (Marchenko et al., 2008, 2010) was used to assess possible changes in permafrost thermal state and active-layer thickness for the entire northern hemisphere permafrost domain. Input parameters to the model include spatial datasets of mean monthly air temperature and precipitation, prescribed vegetation, and thermal properties of the multi-layered (500 m in depth) soil column and water content (which are specific to each vegetation and soil class, and to geographical location). An ensemble of five IPCC GCMs was used for climate forcing, most of which were among the best-performing models for the Arctic and sub-Arctic (see Chapter 3): ECHAM5, GFDL21, CCSM, HADcm and CCCMA (based on the A1B emissions scenario). The model output has been scaled to a horizontal resolution to 0.5° latitude and longitude. According to this specific climate scenario, projections of future changes in permafrost suggest that by the end of the 21st century, late-Holocene permafrost in the Northern Hemisphere may be actively thawing at the southern boundary of the permafrost domain and Late Pleistocene permafrost could start to thaw at some locations (Figure 5.10). However, there is some cold bias for this particular climate forcing in comparison with the CRU2 dataset, especially for the northwestern Eurasia area (the Russian North). These projections were therefore adjusted in compliance with the CRU2 dataset (Mitchell and Jones,

2005) using the difference between observed, interpolated, and projected climate data during the 1981 to 2000 period.

5.2.3.2. Regional permafrost projections

5.2.3.2.1. Alaska

New and detailed results from modeled permafrost dynamics in Alaska have become available since the Arctic Climate Impact Assessment (ACIA, 2005). The GIPL2 model has been used to investigate how observed and projected changes in climatic parameters (mainly air temperature and precipitation) influence permafrost dynamics in Alaska (Marchenko et al., 2008). The model was validated using precise ground temperature measurements from shallow boreholes across Alaska (Romanovsky and Osterkamp, 1997).

The model was run at a resolution of 0.5° latitude × 0.5° longitude spanning the entire Alaskan permafrost domain for the 1900 to 2100 time interval. The CRU2 dataset (Mitchell and Jones, 2005) was used as climate forcing for the period 1900 to 2000. The future climate scenario was derived from the MIT-2D integrated global system model developed at the Massachusetts Institute of Technology (Sokolov and Stone, 1998), using a gradual doubling of atmospheric CO_2 concentration that corresponds to the IPCC A1B emissions scenario.

Ground temperatures were compared at depths of 2, 5, and 20 m for three snapshots in time: 2000, 2050 and 2100 (Marchenko et al., 2008: Figure 5.11). In comparison with present-day conditions, model results show the greatest changes in temperature for the 2050 and 2100 time slices will occur at 2 m depth (Figures 11a, b, c). By the end of the 21st century, mean annual ground temperatures at 2 m depth could be above 0 °C everywhere southward of 66 °N except for small patches at high altitude in the Alaska Range and Wrangell Mountains (Figure 5.11c). An area of approximately 850 000 km^2 (about 57% of the total area of Alaska) could experience widespread permafrost degradation and by 2100 could contain areas in which permafrost will completely disappear in addition to areas where thawing is still ongoing. (The term 'thawing permafrost' indicates a depressed permafrost table and a residual thawed layer – 'talik' – between the seasonally frozen layer and permafrost table that exists throughout the year.)

According to the model results, the extent of the area with mean annual ground temperatures at 5 m depth above 0 °C in 2000 is approximately 125 000 km^2 but will be around 659 000 km^2 (about of 45% of the total area of Alaska) by 2100 and extend into the Alaskan interior (Figure 11f). While permafrost temperatures at 20 m depth could change significantly (although remaining below 0 °C), model results suggest the area with mean annual ground temperatures above 0 °C at this depth will increase by less than 100 000 km^2 between 2000 and 2100 (Figures 11g, h, i). Projected changes in permafrost temperature are more pronounced in areas that currently have colder permafrost than those where temperature is presently close to 0 °C or in peatlands with a deep organic layer.

The simulated mean values of active-layer thickness for the whole Alaskan permafrost domain are 0.78, 1.33 and 2.4 m for 2000, 2050, and 2100, respectively. The projected area of thawing permafrost (for areas where the permafrost table is located deeper than 3 m) also increased (Marchenko et al., 2008), from 65 000 km^2 in 2000, to 240 000 km^2 by 2050, to 720 000 km^2 by 2100 (Figure 5.12).

5.2.3.2.2. Canada

Since the Arctic Climate Impact Assessment (ACIA, 2005), several studies have projected permafrost conditions over the 21st century for all or part of Canada. Zhang et al. (2008a,b) simulated permafrost distribution for the whole Canadian landmass at a resolution of 0.5° latitude × 0.5° longitude using a dynamic model driven by six climate change scenarios. Simulated change in near-surface ground temperature was much greater than for deeper layers and the ground thermal regime was in strong

disequilibrium with the climate conditions projected for the 21st century (Zhang et al., 2008a). However, thawing of the upper part of permafrost does not imply complete loss of permafrost because changes in deeper ground temperature lag behind those at the surface, as is clear from long-term measurements in deep boreholes (e.g., Osterkamp, 2005).

The upper 2 to 3 m of permafrost is projected to thaw over 16% to 20% of the area currently underlain by permafrost in Canada by the end of the 21st century (Zhang et al., 2008a,b). This is a much smaller estimate than earlier estimates from equilibrium projections, for example, Kettles et al. (1997) had projected a 43% reduction. Permafrost thaw would continue beyond the 21st century even if air temperature stops increasing because of the disequilibrium between deeper ground thermal regime and the current climatic conditions. The model results suggest that active-layer thickness would increase significantly, by 0.3 to 0.7 m (or 41% to 104%), and supra-permafrost taliks would appear in many southern regions. About 9% to 22% of land with permafrost in Canada would contain taliks by the end of the 21st century, and the permafrost table would be 1.9 to 5.0 m deeper, mainly due to the formation of taliks. Other modeling studies also suggested the formation of taliks with climate warming (e.g., Delisle, 2007; Duchesne et al., 2008), and taliks have already been observed in some regions (Oberman, 2008). Permafrost thaw would also occur from the bottom up in some southern regions (as observed by Johansson et al., 2008 in northern Sweden). However, concurrent changes in snow conditions (depth, annual duration, timing) with climate change could somewhat mitigate permafrost thaw (Zhang et al., 2008c; Taylor et al., 2006).

Changes in ground thermal regime in northeastern Canada from the present (1961 to1990) to the future (2041 to 2070) were simulated by Sushama et al. (2006) at a 45 × 45 km resolution using a one-dimensional heat conduction model (Goodrich, 1982) driven by surface temperature and snow depth from the Canadian Regional Climate Model (Laprise et al., 2003). Sushama and co-workers projected significant warming in near-surface ground temperature and an increase in active-layer thickness of more than 50% over most of the continuous permafrost region in northeast Canada. Duchesne et al. (2008) simulated permafrost conditions for the Mackenzie River Valley at a higher spatial resolution (1 × 1 km) and results suggested that permafrost degradation would occur mainly through deepening of active-layer thickness and the development of taliks.

Woo et al. (2007, 2008) used a probabilistic approach to project the range in thaw depth that would occur in response to climate warming over the 21st century in the Mackenzie Valley region, an area where recent increases in air temperature have been the greatest. The results for the boreal and tundra environments indicate that thaw depth will increase 15% to 40% over the 21st century in response to warming, with smaller increases occurring where a thick organic layer is present (Woo et al., 2007).

There has been limited exploration of climate change impacts on mountain permafrost within the polar regions. The distribution of permafrost in mountainous areas is particularly complex (e.g., Lewkowicz and Ednie, 2004). Within the Yukon Territory of Canada the existence of mountain and latitudinal permafrost is difficult to delineate as permafrost may exist at sea level in this region. Advances have been made in recent years to model and map the current distribution of mountain permafrost over this extensive region at high resolution (Lewkowicz and Ednie, 2004; Bonnaventure and Lewkowicz, 2008; Lewkowicz and Bonnaventure, 2008). An empirical stochastic model was developed by Haeberli (1973) based on topographic parameters and snow cover (known as the Basal Temperature of Snow (BTS) method). This work has now been extended to include the effects of climate change and to characterize the sensitivity of mountain permafrost within three basins in this region to warming and to project future equilibrium permafrost distribution (Bonnaventure and Lewkowicz, 2010). Increases in mean annual air temperature of up to 5 °C were considered in the analysis and results suggest climate change could have significant impacts on permafrost extent and spatial distribution in the southern Yukon area indicating loss of permafrost in two of the three basins considered.

5.2.3.2.3. The Nordic Area

Lowland permafrost is rare in Fennoscandia but exists in the northern parts of Norway, Finland and Sweden mainly as palsa mires. The spatial distribution of palsa mires in sub-Arctic Fennoscandia is highly correlated with air temperature and precipitation patterns (Luoto et al., 2004a). Fronzek et al. (2006) developed a correlative model using climate-envelope techniques to simulate areas climatically suitable for palsa mires as a function of monthly temperature and precipitation. Results indicated that it was likely (>66%) that palsa mires would disappear completely by the end of the 21st century under scenarios of medium (A1B scenario) and moderately high (A2 scenario) greenhouse gas emissions (Fronzek et al., 2010). For a low emissions (B1) scenario, it was more likely than not (>50%) that conditions over a small fraction of the current palsa distribution could sustain palsas until the end of the 21st century.

Most of Fennoscandia has alpine topography. However, overall changes in permafrost distribution have to date only been modeled on the basis of air temperature (Figure 5.13) (Harris et al., 2009). In Svalbard, Isaksen et al. (2007a) demonstrated recent warming and showed that the extreme year of 2006 lies at the upper percentile of projected climate scenarios for the area. A one-dimensional heat flow model was calibrated with high accuracy, and forced by a suite of empirically downscaled IPCC AR4 GCMs (Benestad, 2007). The model results projected a warming of permafrost, and the development of a significantly thicker active layer and taliks both in bedrock and sedimentary material by 2100 (Etzelmüller et al., 2011). For Iceland, future climate change impacts on permafrost are represented by a sensitivity study for increased temperatures and snow cover based on four shallow boreholes. As in the Scandinavian mountains, Iceland experiences relatively warm permafrost close to the thawing point due to high geothermal activity, a maritime climate and associated snow cover. Even small changes in temperature and snow cover almost immediately result in degradation of permafrost in the mountain settings (Etzelmüller et al., 2007, 2008; Farbrot et al., 2007). High geothermal gradients in Iceland restrict the thickness of permafrost, and facilitate a faster response to climate change than in Svalbard or the Scandinavian mountains.

5.2.3.2.4. Russia

New projections of future changes in permafrost in the Russian European North (Marchenko et al., 2009, 2010; Stendel et al., 2010) have a horizontal resolution of only 4 km. Averaged data for the period 2030 to 2049 suggest significant change (Figure 5.14) and project that by the end of the 21st century, late-Holocene permafrost may be actively thawing at all locations within the Pechora River catchment. Also, some Late Pleistocene permafrost to the north of the Pechora River catchment could start to thaw in some locations. The model results also show how different types of ecosystem affect the stability and thermal state of permafrost.

Oberman and Liygin (2009) used a statistical model linking air and ground temperatures to construct regional permafrost projections for 2020. The model was validated using permafrost monitoring data from a large number of observational sites and boreholes in the north European part of Russia. The forcing climate scenario was based on an extrapolation of air temperature trends calculated for the 1970 to 2005 period from observations at weather stations.

Results project more pronounced changes in permafrost in the eastern part of the study region by 2020 (Figure 5.15) with ground temperature increasing by 0.6 to 1.0 °C. This is consistent with the projected pattern changes in air temperature. It is also consistent with the dominance of coarse grained soils and bedrocks that have relatively high thermal conductivity and are therefore more responsive to climatic warming than other soil types.

5.2.3.2.5. Model inter-comparison challenges

Results of predictive permafrost models presented in this section reflect a wide range of geographical scales, from the pan-Arctic to the local scale. Although the models differ conceptually in complexity and

level of detail, most may be applied at a range of scales if the forcing data are available. For a given level of data availability, there is an optimal range in model complexity beyond which there is a reduction in predictive performance. There are either too many parameters required by the model and an inadequate amount of empirical information to uniquely identify these parameters, or the model is too simple to account for important processes and to fully exploit available empirical information. The interplay between data availability and complexity of computational algorithm is a key constraint in seamless predictive permafrost modeling and model intercomparison.

5.3. Impacts of changes in permafrost distribution and extent

- Recent analyses indicate contrasting changes in hydrology in permafrost regions over past decades: landscape dryness is increasing in the boreal forest, particularly in areas of discontinuous permafrost, whereas some sub-Arctic areas are experiencing waterlogging when permafrost thaws.

- Biodiversity and ecosystem processes on land and in freshwaters are being affected by changes in hydrology. Recent studies show that thawing of ice-rich permafrost is leading to the draining of wetlands resulting in a loss of habitat in some areas whereas in others, thawing permafrost is leading to impeded drainage and a shift in biodiversity to wetland vegetation. Thaw slumping may affect ecosystems sooner than air warming alone.

- Since the Arctic Climate Impact Assessment was published in 2005, new research has demonstrated the viability and diversity of organisms preserved in ancient permafrost.

- New measurements and estimates of soil carbon pools in permafrost indicate that there is more soil carbon in permafrost than previously thought. This carbon is likely to play a more important role in feedbacks to climate during permafrost thawing than earlier calculations suggested.

- Very high emissions of the powerful greenhouse gas nitrous oxide have recently been discovered from terrestrial permafrost in Arctic regions. Although the importance of the process cannot be generalized across the Arctic, the strong radiative forcing potential of the gas suggests important potential contributions to climate forcing.

- High concentrations of subsea CH_4 from the Laptev Sea have recently been recorded throughout the water column and in the atmosphere. Although more measurements are needed from other seas with different permafrost shelf conditions, the current observations reinforce concerns about a major feedback to the climate system from thawing and destabilization of subsea permafrost.

- Recent studies at some locations in the Arctic show increases in thermokarst development which are specifically important where ice-wedges degrade or palsas (peat mounds with a frozen core) decline. Permafrost degradation is also found likely to increase slope instability by increasing the amount of rockfalls and rockslides, and to increase rates of rock glacier movement. In areas with coastal permafrost, warming of permafrost may also increase coastal erosion.

- Although there has been limited scientific evidence to date that observed climate warming has been the direct cause of failure of engineering structures on permafrost, new studies are suggesting that this effect is sometimes important.

5.3.1. Hydrological processes and responses to thawing permafrost

Hydrological processes considered in this chapter concern the direct impacts of thawing permafrost at the landscape level, but not consequences for river flow as these are addressed in Chapter 6.

5.3.1.1. Interactions between hydrological processes and permafrost

Important hydrological processes in the terrestrial Arctic include snow and rainfall precipitation, redistribution of snow by strong wind, snowmelt, damming of snowmelt runoff by the snowpack, surface storage in small depressions and ponds, infiltration and storage in the substrate (e.g., soil), evaporation, transpiration, and runoff (Woo, 1986). The primary control on local hydrological processes is dictated by the presence or absence of permafrost, but they are also influenced by the thickness of the active layer and by the total thickness of the underlying permafrost. As permafrost becomes thinner or decreases in areal extent, the interaction of surface and sub-permafrost ground water processes becomes more important.

The inability of soil moisture to infiltrate deeper groundwater zones due to ice-rich permafrost, maintains very wet soils across the Arctic region in general, and particularly in flat areas and where the frozen layer is near the ground surface (Kane et al., 2004). Although not common, drought conditions have been observed in Arctic Alaska (Kane et al., 2008). These very dry conditions accompanied unusual environmental responses, which included low or no flow in small rivers and the very uncommon occurrence of tundra fire due to the exceptionally dry surface organic layer together with an increased frequency of lightning. These areas have relatively high evapotranspiration and sensible heat transfer, and a low conductive heat transfer due to the insulative properties of thick organic soils. In contrast, in the slightly warmer regions of the sub-Arctic, the thinner or discontinuous permafrost and permafrost-free areas often have quite dry surface soils as soil moisture infiltration is not as restricted. Whether the surface soils are wet or dry greatly affects ecosystem dynamics, fire frequency, the surface energy budget through latent and sensible heat fluxes, and biogeochemical cycling (including trace gas feedback to the climate system).

Short-term active-layer dynamics and longer-term permafrost dynamics affect water storage and run-off. As the climate warms, the active layer thickens, and there is greater storage capacity for soil moisture and greater lags and delays introduced into the hydrological response times to precipitation. When the permafrost is very close to the ground surface, stream and river discharge peaks are higher and base flow is lower. As permafrost becomes thinner, there can be more connections between surface and subsurface water. As permafrost extent decreases, there will be more infiltration to groundwater and the timing of stream runoff will change. Permafrost degradation will result in increased winter stream flows, decreased summer peak flows, and changes in stream water chemistry. Thawing permafrost is also associated with several geomorphological processes, some of which (e.g., talik formation and thermokarst, as described in the following section) are hydrologically important as they permit drainage of soil moisture throughout the year, promoting drying of the surface.

5.3.1.2. Recent changes in landscape hydrology due to permafrost thaw

A number of recent analyses indicate that landscape dryness has increased in some areas of the Arctic whereas in others it has decreased. Landscape dryness has increased over the past several decades in the boreal forest, particularly in areas of discontinuous permafrost (Yoshikawa and Hinzman, 2003; Smith et al., 2005b; Riordan et al., 2006; Burn et al., 2007; Goetz et al., 2007; Figure 5.16). In the southern discontinuous permafrost zone of western Siberia, Smith et al. (2005b) reported an 11% to 13% decrease in lake area between 1973 and 1998. Also, in sub-Arctic areas in Alaska, historical aerial photographs have been used to document a decrease in waterbodies due to subsurface drainage or tapping of thermokarst lakes and other factors (Yoshikawa and Hinzman, 2003; Riordan et al., 2006). Drainage of thaw lakes in the western Canadian Arctic between Inuvik in the Northwest Territories and the Beaufort Sea may be related to melting of channels through ice-rich permafrost. Retrogressive thaw slumps could also trigger lake drainage and Lantz and Kokelj (2008) have noted that slump activity in the region has increased since 1950, probably due to warming temperatures. Slumping may also be promoted through the expansion of taliks (which could increase with warmer lake temperatures) adjacent to the lakes and the increase in ground temperature that results from the disturbance of the organic mat that accompanies the initial slump (Kokelj et al., 2009).

Although warmer conditions associated with climate warming might be expected to result in increased permafrost thawing and increased lake drainage, there appears to be a complex interaction between climate, ice-wedge cracking, retrogressive thaw slumps and lake hydrology (Marsh et al., 2009). As a result, in contrast to drying of thermokarst lakes, some areas are currently experiencing waterlogging when permafrost thaws, such as in sub-Arctic Sweden (e.g., Johansson et al., 2006; Figure 5.17). Using Landsat satellite imagery Smith et al. (2005b) reported a 12% increase in lake area in the continuous permafrost zone in western Siberia between 1973 and 1998. Also, Walter et al. (2006) reported a 15% increase in lake area near Cherskii, Russia. In the Altai Mountains in Siberia, lake area increased by 52% between 1952 and 2007, with the greatest increase after 1972 (Borodavko, 2009). In northwest Canada, Plug et al. (2008) found both large increases and large decreases in lake area over time, and found also that lake area was strongly dependent on cumulative precipitation in the preceding 12-month period.

5.3.2 Geomorphological processes

Geomorphological processes change the land surface affecting human habitability, natural resources, and ecosystem function. Knowledge of the interactions among the different geomorphological processes and permafrost dynamics is crucial for facilitating adaptation to changes and impacts of these processes. Although some geomorphological responses to thawing permafrost can potentially create local hazards, the importance of this field of study is only recently being recognized and significant attempts are being made to monitor the different geomorphological processes addressed here.

5.3.2.1. Thermokarst

5.3.2.1.1. Thermokarst landforms and processes

Thermokarst is a general term relating to the melt of ground ice and the very irregular surface landforms produced through permafrost degradation. Thawing can be controlled by natural or human-induced geomorphical, vegetational or climatic processes (French, 2007). These processes are capable of rapid and extensive modification of the landscape.

Permafrost degrades in a continuum from rising temperatures in frozen ground (which increase the unfrozen water content and reduce the load-bearing strength of the ground) to complete thawing of ice-rich ground (which causes the surface to subside and creates depressions in the ground, termed 'thermokarst'). The types of thermokarst landforms and their ecological implications are extremely variable depending on climate, topography, soil texture, hydrology, amounts and types of ground ice, and heat transfer mechanisms (Figure 5.18). These landforms and processes generally occur at mean annual air temperatures as high as +2 °C, and as low as -20 °C in the High Arctic, and vary greatly in completeness of degradation. Thawing can occur downward from expansion of the active layer, laterally from water and radiation, internally from groundwater intrusion, and upward from the bottom due to geothermal heat flux.

Lateral degradation from an expanding deep thermokarst lake in continuous permafrost, leads to talik formation and intermediate degradation, but complete degradation of permafrost is rare because of the thickness of the permafrost. In contrast, lateral degradation from a thermokarst lake in the discontinuous zone quickly leads to complete degradation because permafrost is thinner. At least 22 thermokarst landforms have been differentiated (Jorgenson and Osterkamp, 2005) and they can be at various stages of degradation and stabilization (Burn and Smith, 1990; Leibman and Kizyakov, 2006; Jorgenson et al., 2008; Veremeeva and Gubin, 2008).

The number and extent of hillslope thermokarst landforms (e.g., active-layer detachment slides and thaw slumps) have been quantified recently using aerial photographs (Lantuit and Pollard, 2008; Lantz and Kokelj, 2008; Goosef et al., 2009). However, this is difficult because thermokarst landforms range in size from small thermokarst pits to large thermokarst lakes, and surface conditions vary from water to different vegetation structures that confound remote sensing. A regional study for Alaska revealed that within the

discontinuous permafrost zone, 5% of the area had thermokarst, 62% had permafrost, 21% was unfrozen with no recent permafrost, and 12% was undetermined (Jorgenson et al., 2008). In the continuous permafrost zone, thermokarst terrain was evident on 13.5% of the area, and 1.5% was unfrozen under deep non-thermokarst lakes. The high prevalence of thermokarst landforms in the colder continuous permafrost region was attributed by Jorgenson and co-workers to the occurrence of massive ice in the form of ice wedges, and to problems differentiating between actively and inactively degrading landforms.

5.3.2.1.2. Changes in thermokarst landforms

Rates of change in thermokarst have frequently been determined using remote sensing of waterbodies because they tend to be spectrally distinct, although detecting trends using only a few periods of moderate-scale Landsat (28 to 80 m) imagery is problematic. Waterbodies have recently both decreased and increased in frequency in the Arctic permafrost regions (see Section 5.3.1.2).

Thaw slumps and active-layer detachment slides have been increasing in the Northwest Territories, Canada (Lewkowicz and Harris, 2005; Lantz and Kokelj, 2008; Lacelle et al., 2010), northwestern Alaska (Gooseff et al., 2009), Russia (Leibman, 1995), and along the Arctic coast (Lantuit and Pollard, 2008). In the southern permafrost region of Canada, where most of the permafrost was formed during the Little Ice Age, 9% of the permafrost has degraded into thermokarst bogs and fens since it was formed (Vitt et al., 2000), and field monitoring indicates that lateral rates of thawing are accelerating (Camill, 2005). Across four sites representing different regions in Alaska, aerial photograph analysis revealed that the total area affected by thermokarst increased 3.5–8% over about 50 years (Jorgenson et al., 2008).

Predicting the geomorphical responses of thermokarst terrain to future climate warming is complicated by the complex interaction of biophysical factors. These create strong negative feedbacks from vegetation-soil processes (see Section 5.3.3.2) that make permafrost more resilient to warming and disturbance, as well as positive feedbacks from thaw settlement and water impoundment that make permafrost more vulnerable to warming (Jorgenson et al., 2010). Vegetation-soil feedbacks can reduce deep soil temperature by 7 °C and help permafrost to persist at mean annual air temperatures of up to +2 °C, while water impounded by thaw settlement can increase ground temperature by 10 °C, and make permafrost vulnerable to thawing at mean annual air temperatures as low as -20 °C (Jorgenson et al., 2010). These feedbacks are larger than the projected climate warming of 3 to 6 °C for much of the Arctic, and greatly complicate a prediction of how permafrost will respond to climate warming.

5.3.2.2. Periglacial landforms and processes

Periglacial environments are defined by mean annual air temperatures below 3 °C (or 5 °C in windy regions) and the most important ecological boundary by which they are delimited is the treeline. Periglacial landforms are controlled largely by the presence and thermal state of permafrost, and these landforms are key indicators of changes in permafrost and are also important in determining landscape stability.

5.3.2.2.1. Ice-wedge polygons

Ice-wedge or tundra polygons are probably the most widespread periglacial landforms in lowland continuous and discontinuous permafrost areas (Figure 5.19). Because their borders are underlined by networks of ice wedges, their presence implies large volumes of ground ice near the surface of the terrain (Harry and Gozdzik, 1988). During warming it is likely that the resulting ground thawing of ice wedges will affect permafrost landscapes significantly, leading to disturbed topography and drainage patterns – a first stage in thermokarst development. However, the impact of warming on polygonal terrain is likely to be very variable in the Arctic, depending on regional climatic regimes and local terrain conditions. The ice-wedge responses in the cold continuous permafrost zone will be less pronounced than responses further south, near the boundary between continuous and discontinuous permafrost. Changes in snowmelt

that affect the water regime, together with summer precipitation have already led to destruction of ice-wedge networks through thermo-erosion (Fortier et al., 2007). For example, cessation of ice-wedge development occurred during the 20th century in the Mackenzie Delta region of northern Canada (Kokelj et al., 2007).

The response of ice wedges to climate change is largely controlled through the basic physics of thermal contraction cracking that governs their initiation and growth, in addition to thermally induced seasonal movements of the active layer and upper permafrost (Mackay, 2000). Thermal contraction cracking occurs in the coldest winter periods, when ground temperature drops to -15 °C at the top of the permafrost. This typically occurs during cold events, when near-surface rapid atmospheric cooling occurs, and air temperature drops 10 to 20 °C (e.g., from -10 °C to -35 °C) over short periods (Christiansen, 2005; Fortier and Allard, 2005). The fast ground cooling then leads to a rate of ground contraction which exceeds the soil strength, leading the soil to crack open. Although the cracks close to some extent when the ground starts to warm in spring, the cracks remain open until snowmelt, at which point water flows in and freezes, forming ice veins that grow into ice wedges over the years.

Even with the climate warming projected for the coming decades, it is expected that meteorological conditions in the coldest winter months will continue to sustain ice-wedge formation, and that ice-wedge polygons will continue to be active Arctic landforms. However, during warming, the active layer in polygon centers and over ice-wedge tops will probably get deeper by several centimetres or tens of centimetres. Thawing of ice-wedge tops will lead to localized thaw settlement producing deeper troughs along the polygon sides. This settlement will affect vegetation, local drainage, and snow cover distribution in winter, leading to changes in cracking activity, but with no immediate major geomorphological change.

Thawing of ice wedges under cold temperatures in northern Alaska typically causes surface degradation only and rarely leads to the intermediate degradation that is characterized by the formation of a talik (Jorgenson et al., 2006). Also in northern Alaska, thermokarst associated with ice-wedge degradation increased from 0.5% of the land area in 1945 to 4.4% of the land area in 2001, based on aerial photograph analysis of small-scale features (Jorgenson et al., 2006).

Ice-wedge polygons are especially sensitive to thermal erosion induced by the flow of running water in the ice-wedge troughs. Excess surface flow in the short summer period eventually erodes the ice-wedges because of the large amount of heat supply from convection. Tunnels and ditches develop along the ice-wedge networks and lead to ground collapse and very significant terrain disturbances (Figure 5.20) (Fortier et al., 2007) as a first stage in thermokarst development. It is possible that, as the frequency of rapid spring melts and rain events in summer increase, thermo-erosional events over polygonal terrain will occur more often. This is most likely to expose to risk some man-made structures built over ice-wedges such as roads, runways and pipelines.

The frequency of frost cracking will be reduced in the southern part of the permafrost zone as winters become warmer and cooling events occur less frequently, or if a thicker snow cover accumulates that enables insulation of the ground from atmospheric cooling. Field experiments have indeed demonstrated that a thick snow cover (more than 60 to 80 cm) locally prevents frost cracking (Kokelj et al., 2007). However, even under a warmer climate, sites cleared of snow will continue to endure thermal cracking, a process that seriously affects man-made structures such as roads and runways (see Section 5.3.5.1.3).

5.3.2.2.2 Pingos

Pingos are ice-cored hills, typically conical in shape, that grow and persist only in a permafrost environment (Mackay, 1998). They range in height from a few metres to several tens of metres (Figure 5.21) and are found in continuous and discontinuous permafrost regions. Pingos are important biodiversity hot spots (Section 5.3.3.1).

The core of massive ice in pingos is mainly formed by water injected under pressure. Two types of pingos are recognized. First, hydraulic, open system pingos, into which the water is injected by artesian pressure, that is, where a hydraulic gradient in the surrounding terrain would drive groundwater to the formation site (e.g., Ross et al., 2007). Second, hydrostatic, closed system pingos, into which the water is concentrated by pore-water expulsion in the soil. This is caused by permafrost aggradation beneath the bottoms of drained lakes that are underlain by saturated sediments. Hydraulic system pingos are less understood than hydrostatic system pingos, which are abundant in flat plains covered by hundreds of lakes in Canada (e.g., Mackay, 1979, 1998).

Observations and a drained lake experiment (Mackay, 1997) have shown that pingos form when a shallow lake suddenly drains due to geomorphological changes, and permafrost aggrades from all directions in the newly exposed sediments of the previously unfrozen lake bottom (a sub-lake talik). Over the years, this builds up the pressures that will concentrate the water into a bulb, pushing the terrain upward over it and creating a mound that will eventually freeze totally, making an ice core. Many tundra lakes drain suddenly when their banks breach due to coastal or fluvial erosion, or when their water flows over ice-wedges and rapidly erodes a drainage outlet by thermo-erosion. Pingos ultimately collapse when cracks on the sides and summits expose the ice core to the warm summer atmosphere; a crater-like depression may develop in the core of the mound, sometimes occupied by a thaw lake; and slumps occur on slopes. Pingo development is a natural cyclical process, with a single cycle taking up to several thousands of years.

The impacts of climate warming on Arctic pingos over the coming decades are difficult to project quantitatively, given the complex interactions between heat exchange and the groundwater under pressure, and local geomorphological controls on different water sources. For hydrostatic system pingos, which still have a lens of pressurized unfrozen water beneath the ice core, a change in their complex internal thermal regime may affect their stability or growth rate. As hydraulic system pingos are controlled by the supply of water from outside the pingo itself, their activity will also largely depend on changes in the surrounding environment. These include changes to glaciers that have an unfrozen zone and can provide the water. However, some controls are non-climatic, for example, springs are often under geological control, and sometimes the activity of geological faults creates feeding channels for the steady water supply needed.

Overall, degradation of collapsing pingos (Figure 5.22) is likely to accelerate and the number of degrading pingos could increase with time, leading to increased thermokarst landscape development. However, continued coastal erosion as well as increased thermo-erosion of ice-wedges are likely to result in the draining of lakes, which will favor permafrost expansion under newly exposed lake beds. Therefore, inception of new pingos is not impossible in regions that will remain cold enough for some time.

5.3.2.2.3. Palsas and peat plateaus

Palsas are peat mounds with a frozen core and located in the southern parts of the permafrost zone (Seppälä, 1988). Declines of palsas, palsa mires and peat plateaus (both extent and abundance) have been observed in North America (Payette et al., 2004; Camill, 2005; Sannel and Kuhry, 2009) and northern Europe (Zuidhoff and Kolstrup, 2000; Luoto and Seppälä, 2003; Akerman and Johansson, 2008; Sannel and Kuhry, 2008) due to a warming climate (see Section 5.2.2). Thermal erosion of palsas and peat plateaus often results in the development of thermokarst lakes (see Section 5.3.2.1).

5.3.2.2.4. Frost weathering

Frost weathering results from freezing and thawing of water within rock or mineral particles (Matsuoka and Murton, 2008) and can lead to rock fractures and slope instability that can create hazards. The 9% volumetric expansion of water as it changes from liquid to solid phase has traditionally been considered the main causal mechanism of frost weathering, although recently the role of ice segregation within certain bedrock lithologies has been recognized as potentially of greater significance (Hallet, 2006). The main difference between the two processes is that volumetric expansion arises from *in situ* freezing of

water, whereas ice segregation involves water migration within freezing or frozen ground (Rempel, 2007). Volumetric expansion occurs at the freezing point of the water occupying pores or cracks and requires a high saturation level (>90%) of the rock. In contrast, ice segregation can occur in unsaturated rock. The presence of capillary and adsorbed water with a freezing point below 0 °C allows unfrozen porewater to migrate through partially frozen rock to supply progressive growth of ice lenses. The resulting rock fracture can occur at temperatures considerably below 0 °C (Walder and Hallet, 1985; Akagawa and Fukuda, 1991; Hallet et al., 1991), depending partly on the pore structure (Matsuoka, 2001a). If ice segregation in bedrock permafrost is widespread, there may be considerable potential for significantly increased rock slope instability associated with rising ground temperatures and thickening active layers in a period of climate warming.

Recent experiments simulating a bedrock active layer above permafrost have shown that moist, porous rock (chalk) behaves remarkably like moist, frost-susceptible soil, with both substrates experiencing ice enrichment and fracture/fissuring of near-surface permafrost (Murton et al., 2006). At the start of the experiments, the rock was unweathered and lacked visible fractures. However, after repeated cycles of active-layer freezing and thawing, fractures filled with segregated ice had formed in the transition zone between the active layer and the permafrost (Figure 5.23) due to downward migration of water in summer and upward advance of freezing at the beginning of winter.

The importance of ice segregation relative to volumetric expansion increases with decreasing thermal gradients and increasing duration of freezing (Powers and Helmuth, 1953; Walder and Hallet, 1985). Rempel et al. (2004) suggested that the maximum possible disjoining pressure is governed by the temperature depression below the bulk-melting point, even in the absence of large temperature gradients, and therefore slow ice segregation in bedrock may be possible at greater depths where the frozen permeability of rock limits the actual amount of heave produced. Thus, over long time-scales, ice segregation may be highly significant in frozen steep bedrock slopes, where the presence of ice-rich fractured bedrock may be critically important in releasing rockfalls and rockslides during climate warming and permafrost degradation (Gruber and Haeberli, 2007).

5.3.2.3. Slope processes and permafrost-related geo-hazards

5.3.2.3.1. Rock glaciers and glacier-permafrost interaction

Rock glaciers are characteristic large-scale flow features of frozen rock material that is the geomorphological result of a cold climate, with conditions for permafrost formation in high-relief regions (Humlum, 1999). They are located at the foot of rock free-faces with a high supply of talus and, when active, typically take the form of 20- to 100-m thick tongue- or lobe-shaped bodies, with cascading frontal slopes standing at the angle of repose (Figure 5.24). Rock glaciers may be as much as several kilometres long, but are typically 200 to 800 m measured parallel to the flow direction (Barsch, 1996). Their surface is covered by coarse (0.2 to 5 m) rock fragments, and displays a 1- to 5-m high curving transverse furrow-and-ridge topography. Active rock glaciers typically flow downslope by permafrost creep of the order of 0.1 to 1 m per year (e.g., Barsch, 1996; Haeberli et al., 2006), that is, they are more sluggish than normal glaciers, but are nevertheless highly efficient agents of coarse debris transport (Humlum, 2000).

The limited recent evidence on the effect of climate change on rock glaciers suggests that when rock glacier permafrost temperature approaches 0 °C, the creep rate may increase (e.g., Leonard et al., 2005; Roer et al., 2005; Krainer and Mostler, 2006; Kääb et al., 2007). Presumably, this is because more liquid water will be present within the ice at higher temperatures.

The existing evidence on rock glacier dynamics is, however, conflicting (Janke, 2005), suggesting the importance of several factors in addition to air temperature. These include changes in rockfall intensity and snow avalanche frequency. Climatic change resulting in increased snow precipitation and increased snow drifting might result in a higher frequency of snow avalanches on leeward slopes. This may result in

increased creep rates of rock glaciers lining the foot of such slopes, as snow avalanches represent an important source of water for ice formation in rock glaciers and for the supply of rock debris to rock glaciers (Humlum et al., 2007). Rock weathering rates of the bedrock in the headwalls above rock glaciers are controlled by temperature and variations in moisture (and ice) content, and climate is thus important for the rock weathering rate and rockfall frequency (Humlum, 1997). Thus, the relation between rock glacier dynamics and climate change is complex, and still only partially understood. It is likely, however, that warming of rock glacier permafrost temperatures to near 0 °C may result in a period of intensified rock glacier creep, promoting increased slope activity.

Observations of fossil rock glaciers in the Faroe Islands (Humlum, 1998) provide empirical evidence for increased rock glacier creep and local instability during periods of warming (in this case the end of the Younger Dryas period). However, the period of increased rock glacier creep ends when the ice bodies have been reduced to a critical thickness.

5.3.2.3.2. Rockfalls

A 'rockfall' is the fall of relatively small (< 10 m^3) fragments of rock debris that are released from bedrock cliffs by weathering, often accumulating to form a talus (Figure 5.25). Rockfall in periglacial environments has been widely attributed to frost wedging, that is, the widening of cracks and joints by ice during freezing and detachment of debris from cliffs during thaw. Rockfalls may also be triggered by stress-release, progressive failure along joints, or build-up of hydrostatic pressure within a rock mass (André, 2003) and some researchers view the role of freeze-thaw in rockfall as relatively trivial.
The timing of rockfalls when freeze-thaw cycles are assumed to be most frequent, particularly during spring thaw (e.g., Rapp, 1960; Luckman, 1976; Coutard and Francou, 1989), remains the main reason for seeing frost wedging as the foremost cause of rockfall in cold environments. In Japan, Matsuoka and Sakai (1999) observed peak rockfall rate 5 to 15 days after melting of the cliff face, when seasonal thaw reached an estimated depth of 1 m, and found that intensive rockfall activity is rarely associated with either diurnal freeze-thaw cycles or precipitation events. Other rockfall inventories, however, emphasize diurnal variations, rather than seasonal trends (Coutard and Francou, 1989) and the importance of precipitation events. As a result it is difficult to predict the climatic controls on rockfall intensities.

Despite the difficulty in predicting climatic controls, thawing permafrost is likely to affect rockfall intensity. It is very likely that bedrock fracturing by ice segregation (Section 5.3.2.2.4) is significant for rockwall stability in regions of mountain permafrost, and for landform development in such areas in general. Presumably, most segregated ice is found within the topmost 1 to 2 m of bedrock permafrost (French, 2007), so climatic change leading to permafrost thaw would be expected to result in initially higher rockfall intensity, followed by a decrease after this critical layer thawed.

5.3.2.3.3. Rockslides and rock avalanches

Rockslides or rock avalanches represent larger parts of steep rock slopes, which slide or fall downslope (Figure 5.26). This movement can occur at the scale of centimetres per year, but can also be very quick, transporting large amounts of rock all the way to the valley bottom in one event. The greatest geohazard is attached to events on rock slopes located above lakes or the sea as these cause a risk of tsunamis. Rockslides and rock avalanches are influenced by many factors, such as earthquakes and water pressure in the rock. Large rockslides are known to have occurred in several parts of the world, particularly during deglaciation, and these were often associated with increased tectonic activity as the crust is isostatically adjusted. Permafrost formation and thaw can also affect rockslides and rock avalanches.

When rockslides start moving, cracks open in the rockslope, and the rate of movement is monitored at sites where there is increased risk of damage to settlements and infrastructure. Permafrost has been identified in some rockslide areas. If permafrost thaws or forms due to climatic changes, rocksliding might increase and rock avalanches could even occur (Dramis et al., 1995). Nearly all the literature on

permafrost in steep rock slopes and its influence on rocksliding is from non-Arctic areas such as the European Alps, where some three-dimensional modeling has been undertaken to help understand the consequences of rising ground temperature (Wegmann et al., 1998; Gruber et al., 2004). To study how permafrost can affect rocksliding and rock avalanches in the Arctic, more ground temperature observations are needed from rockslide areas in potential permafrost areas. Even if the risk from increasing rockslides and rock avalanches could be important to settlements and infrastructure, projection is not yet possible.

5.3.2.3.4. Solifluction

Solifluction is defined as the slow downslope movement of the active layer over permafrost, or over the seasonally frozen layer in other periglacial, non-permafrost environments. Solifluction can take place on slopes in an undifferentiated fashion or may give rise to landforms that include stripes, steps, lobes and sheets. The process is a combination of frost creep and gelifluction and typical surface velocities are 1 to 9 cm/y (Matsuoka, 2001b). Velocity profiles with depth are concave downslope in areas of warm permafrost and seasonally frozen ground in the discontinuous zone, where autumn freezing is from the surface downward. Convex downslope velocity profiles develop in areas of cold permafrost where freezing is also from the permafrost table upward. The environmental boundary between the two has not been delineated, but is thought to correspond to the boundary of the zone of continuous permafrost. Recent advances in solifluction studies have taken place through a combination of field monitoring (Harris et al., 2007) and laboratory simulation (Harris et al., 2008).

The impact of climate change on solifluction is inferred from understanding of processes and observations during warm summers, and is linked to the distribution of ice within the active and transient layers. A warm summer in an area with two-sided freezing (i.e. freezing from the ground surface downward and from the permafrost surface upward) will result in enhanced late-summer movements, as the base of the active layer and the top part of the transient layer thaw relatively quickly, inducing high porewater pressures. This is similar to what happens in a normal summer, but the depth and amount of movement will usually be greater (Lewkowicz and Clarke, 1998). In an area with one-sided freezing (i.e. from the ground surface downward), most movement normally takes place during and immediately following snowmelt and there may be little or no movement later on (Kinnard and Lewkowicz, 2005). However, a warm summer in which the transient layer thaws will result in movements similar to those in an area of two-sided freezing.

Projection of these differences into the future has not been done and will depend on the rate of ground ice replenishment in the transient layer versus the frequency at which the active layer deepens. Overall, it can be expected that rates of solifluction are likely to increase in the warmer and/or wetter climates that are expected in the Arctic in future, providing permafrost continues to persist.

These enhanced movements may have several impacts. They could increase sediment delivery from slopes to streams and lakes, which could in turn affect aquatic ecosystems. Solifluction can also affect shallow foundations of built structures because of differential movements. However, most structures on permafrost are constructed with foundations that extend well below the active layer, but deepening active layers produced by a warmer climate may allow deeper movements, which could cause problems in the future.

5.3.2.4. Permafrost-glacier interactions

Permafrost-glacier interactions are important to consider when projecting the impacts of changes in glaciers on permafrost. Receding glaciers expose new land to permafrost formation, if the deglaciated areas experience enough ground cooling. Land uplift due to isostatic adjustment as glaciers recede also creates possibilities for permafrost to form. Furthermore, stagnant dead glacier ice (e.g., in ice-cored moraines which are buried below a significant amount of debris) might be preserved if located in a permafrost environment with a debris cover thicker than the active layer. Thermokarst landforms can also

develop from thawing of ice-cored moraines, but such a landscape is traditionally considered as still undergoing areal deglaciation, and thus typically characterized as glacial. In addition, as glaciers become thinner, their thermal characteristics might change and they could become 'cold-based' glaciers, thus increasing the potential for permafrost to form beneath the glacier. Glacier-derived rock glaciers are an example of direct interaction between glacial deposits and permafrost.

5.3.2.5. Coastal and marine processes

Sea ice and ground ice both limit and enhance erosion processes, especially as more than 65% of the Arctic coastline is composed of unconsolidated material (Lantuit et al., 2011), and much of which is ice-bearing or ice-rich. A substantial amount of the Arctic coastline is ice-bound for most of the year, and is thus protected from erosion and deposition. However, despite shallow shelf water, a high ground-ice content and the generally fine-grained unconsolidated material found in lowland sedimentary coasts, render such coasts sensitive to waves and storm surges during the short summer period. These coasts therefore experience relatively high annual erosion rates. Bedrock coastlines are not generally subject to any significant erosion or deposition (Figure 5.27).

In the nearshore zone, where the water depth is less than the maximum ice thickness, bottom-fast ice develops (Figure 5.1). Bottom-fast ice can result in simultaneous preservation and thawing of the permafrost. The high thermal conductivity of the ice creates a thermal diode, shunting heat out of the sediment during the winter months. In some locations where the sediment is already thawed, this creates a frozen active layer beneath the seabed. On the other hand, salt exclusion from the forming sea ice creates highly concentrated brines under pressure (Overduin et al., 2008), sufficiently saline to be unfrozen at temperatures as cold as -10 °C. These brines penetrate the upper sediment and thaw the pore ice under cryotic conditions. In general, subsea permafrost in the bottom-fast ice zone is quasi-stable. Once the water depth exceeds the maximum ice thickness, salt water penetrates into the sediment, and heat transfer from bottom water warms the permafrost. Both processes result in phase change in the sediment pore space, although the sediment remains cryotic and assumes an almost uniform temperature profile near the freezing point determined by its salinity. Submarine permafrost is also subject to degradation from below due to geothermal heat flux (Lachenbruch, 1957).

Most of the significant drivers for coastal erosion point toward increased erosion. These include reductions in sea-ice thickness, extent and duration, which over the past decades have exceeded most forecast scenarios. A resulting increase in open water in the coastal zone brings increased fetch for wave action impinging on the coast, increased heat exchange with the shallow shelf sea water during the summer months, and changes in shelf salinity. Many studies have suggested that the frequency of storm events is increasing (e.g., Are et al., 2008) with a resultant increase in sea-surface height at the coastline due to storm surges. Over fifty years of tide gauge data have now been accumulated for parts of the Siberian Arctic coast, showing a sea-level rise of 2.5 mm/y (Proshutinsky et al., 2007; see also Chapter [sea level rise]). The sedimentary coastal bluffs are sensitive to increased air temperature, which increases thermo-abrasion and thermo-denudation of the coast. Increased ground heat flux thaws ground ice at the top of the permafrost, leading to subsidence and increasing sea-level rate rise relative to land surface elevation. These physical drivers will probably all lead to a system-wide increase in sediment flux in the Arctic, which has already been observed from satellite (Pozdnyakov et al., 2005) in the nearshore zone.

5.3.3. Ecological processes

Permafrost interacts closely with biodiversity and ecosystem processes in the Arctic, and changes in permafrost regimes will affect Arctic ecology. Permafrost moderates soil temperature, drainage, nutrient availability, micro- and mesorelief, rooting depth and plant stability, and disturbance regimes for those species (particularly plants) that live on permafrost. It also provides a total environment for those species that live within permafrost and preserves life from the past. In turn, vegetation moderates ground surface temperature and facilitates disturbance regimes, such as forest fires, that affect permafrost. Understanding

of the interactions between permafrost and Arctic ecosystems has advanced since the Arctic Climate Impact Assessment and the Fourth Assessment of the Intergovernmental Panel on Climate Change and two particularly important issues have emerged. First, the uncertainty of the implications of permafrost thaw for soil water content, subsequent ecosystem development, and greenhouse gas emissions. Second, the preservation and activity of life in permafrost.

5.3.3.1. Life on permafrost – impacts of thawing permafrost on biodiversity

No species are dependent on permafrost (although some ancient microbes might be preserved in it) and at the circumpolar scale, no ecosystems are restricted by the presence of permafrost: both tundra ecosystems and boreal forests can be underlain by permafrost or can occur where it is absent. However, at the landscape scale, the presence of permafrost strongly influences plant species composition (Camill, 1999) and restricts the types of plant that can grow in terrestrial ecosystems. The presence of permafrost affects drainage and determines the presence or absence of fundamental ecosystem types and their species biodiversity: lakes, wetlands, meadows or heaths. Also, shallow active layers restrict the diversity and productivity of trees while cold soils that limit decomposition and nutrient cycling favor vegetation with a low diversity of plant species, and particularly mosses (Van der Wal and Brooker, 2004) that can reduce tree seedling establishment (Camill et al., 2010). At the small scale, active soil movement provides microsites for the establishment of seedlings while preventing plants with rhizomes from establishing (Jonasson, 1986; Jonasson and Callaghan, 1992). Permafrost features that result in soil uplift and drainage (e.g., pingos and baidgerahks) create microsites favorable for biodiversity hot-spots that support, for example, diverse plants and also animals such as snowy owl, as well as lemming nests, wolf dens, and Arctic fox earths (Walker, 1995). Although Arctic vegetation has a simple vertical structure, there is a complex horizontal structure of repeated patterns associated with patterned ground (Matveyeva and Chernov, 2000).

Impacts of thawing permafrost on biodiversity are likely to be numerous and diverse. In many areas of the Arctic, thawing permafrost is leading to the disappearance of lakes and ponds (Section 5.3.3.1) and this will affect biodiversity along the gradient: freshwater ecosystems → wetlands → meadows → heaths. In the polar deserts of the High Arctic, the maintenance of a high water table is critical for the existence of patchy wetlands, which provide hydrological and ecological conditions important to plants, insects, birds and rodents (Woo and Young, 1998). A high water table is maintained by a shallow active layer that restricts drainage. Increase in active-layer thickness resulting from warming of the ground that may accompany climate change will improve drainage and lower the water table. Slumping and erosion that may accompany thawing of ice-rich permafrost can lead to the draining of these wetlands, resulting in a loss of wildlife habitat (Woo and Young, 2006; Woo et al., 2006a). In other areas, thawing permafrost is impeding drainage (Smith et al., 2005b; Johansson et al., 2006) leading to a shift in biodiversity from former dwarf shrub vegetation of dry organic soil to tall shrubs of wetter soils and eventually wetland sedges (Malmer et al., 2005) and bog mosses. In the boreal peatlands of Manitoba, Canada, a plant succession from black spruce (*Picea mariana*) trees to *Sphagnum* moss during permafrost thaw resulted in a doubling of peat accumulation (Camill et al., 2001).

Thawing permafrost is associated with gradual or episodic disturbance (slumping) of the land surface that can affect large areas (Section 5.3.2), with various impacts on biodiversity. It damages existing vegetation, particularly forests when trees become unstable and fall (Figure 5.28) while disturbed organic soils and deeper groundwater systems can occur that yield greater transport of cations, and dissolved and particulate organic carbon, thereby creating favorable conditions for the establishment of spruce trees as well as various shrubs (Lloyd et al., 2003). Slightly drier soils along thermokarst banks promote the introduction of woody species as compared to adjacent tundra. Such processes may accompany a northward expansion of the treeline. Disturbance can also open new niches for the establishment of an enriched flora. Lantz et al. (2009) found that in the Mackenzie Delta region of northern Canada, thaw slumping disturbance could provide opportunities for rapid colonization of species beyond their present range and that the disturbed sites may act as highly productive seed sources within the larger undisturbed terrain. However,

disturbance can also facilitate the establishment of invasive species of detriment to conservation interests and land use. Thaw slumping impacts on Low-Arctic terrestrial ecosystems may be more immediate than the response of the ecosystems to air warming alone (Lantz et al., 2009).

The biodiversity of animal species will be affected by the type of land surface (habitat) resulting from thawing permafrost. For the extreme scenario where thawing permafrost leads to increased drainage and soil drying, the animals associated with freshwaters (e.g., fish and waterbirds) could be replaced by mammalian herbivores and ground-nesting birds etc. However, the opposite scenario applies to the sub-Arctic palsa mires that occur in northern Fennoscandia, Canada, and Siberia. These are heterogeneous and biologically rich environments resulting from the gradients in the water table and the nutrient-rich versus nutrient-poor microhabitats. The palsa mires are projected to almost disappear from northern Fennoscandia by 2050 (Section 5.2.3.2), which would affect biodiversity at the global scale because they are important breeding habitat for birds (CAFF, 2001). This is due to the heterogeneous nature of the environment and to the existence of shallow water that provides abundant food (insects). The disappearance of palsa mires would therefore cause a reduction in populations of migrating birds such as waders that occur exclusively in areas with palsa mires (Luoto et al., 2004b).

Permafrost thaw also leads to rapid changes in the freshwater environment that are likely to affect biodiversity in the shorter term. Kokelj et al. (2008) and Thompson et al. (2008) observed elevated levels of conductivity (reflecting changes in lake chemistry) in small lakes in the Mackenzie Delta region of northern Canada affected by recent (after 1950 - Lantz and Kokelj, 2008) active thermokarst slumping relative to undisturbed lakes. Sloughing of banks (slumping) enhances thawing and erosion, while exposing buried organic material from which soluble material is released (Kokelj and Burn, 2005) and transported by runoff. However, dissolved organic matter was observed to be lower in lakes affected by thermokarst activity (Kokelj et al., 2005). Permafrost disturbance will be an important factor influencing tundra lake chemistry as warming continues in the Arctic and, as freshwater biodiversity is strongly related to water chemistry, changes in biodiversity can be expected. Some of these changes may affect species of local and commercial value (e.g., various fish species).

5.3.3.2. Life on permafrost – impacts of changes in biodiversity on permafrost

Biodiversity (particularly of vegetation) affects permafrost in four fundamental ways: (i) by directly insulating and protecting permafrost; (ii) by indirectly insulating permafrost through its capacity to trap snow; (iii) by affecting land surface albedo and thereby modifying soil temperature (see Section 5.3.4.1); and (iv) creating the conditions for fire that can result in permafrost warming and thaw. In the case of fire, this would occur in the short term through the direct effect of the fire and in the long term because of the effect on surface albedo and vegetation insulating capacity.

Vegetation associated with areas of drier peat or organic soil, insulates the active layer and protects the permafrost (Woo et al., 2006b, 2007; Yi et al., 2007; Smith et al., 2009a). Vegetation-soil feedbacks reduce deep soil temperature by 7 °C and help permafrost to persist at mean annual air temperatures of up to +2 °C (Jorgenson et al., 2010). This has led to 'ecosystem-protected permafrost' such as that in sub-Arctic palsa mires where the permafrost is sporadic in the lowlands. While disturbance to the vegetation could result in permafrost thaw, undisturbed vegetation could allow patchy permafrost to persist even during climate warming (Shur and Jorgensen, 2007).

Snow-vegetation interactions are complex (see Chapter 4). Vegetation, particularly shrubs, can trap snow and cause an increase both in snow depth and snow cover duration. Studies have shown that snow cover insulates the soil. For example, experimental accumulation of snow increased soil temperature over winter by 0.5 to 9 °C (15 cm soil depth) in the sub-Arctic (Seppälä, 2003; Dorrepaal et al., 2004) and by 6 to 15 °C at 50 cm depth in Alaska, depending on the height of the snow-accumulating fence (Walker et al., 1999; Hinkel and Hurd, 2006). Increased snow depth will therefore disconnect permafrost from the low winter temperatures that protect it and will lead to increased thaw (Section 5.2.3.2).

Vegetation has a fundamental impact on surface albedo. Various changes in biodiversity will decrease albedo and increase the positive albedo feedback to the atmosphere (i.e., warming). These include more shrubs in the tundra (Silapaswan et al., 2001; Sturm et al., 2001; Euskirchen et al., 2009), expansion of boreal forest into regions now occupied by tundra (Chapin et al., 2005), and the replacement of summer-green conifers (larch) and other deciduous treeline tree species by evergreen conifers such as pine and spruce (Kharuk et al., 2005; Wolf et al., 2008) (see Section 5.3.4.1 for the feedback effects). Feedbacks to climate from changing biodiversity also include sequestration of carbon from the air, which will help reduce air warming, and increased evapotranspiration that in turn leads to local cooling (see Section 5.3.4.1 for feedback effects).

A major impact of changing vegetation on permafrost owing to general climate warming is likely to be an increased frequency in fire due to drier conditions in existing tundra and forest areas. Also, as projected changes in vegetation (Callaghan et al., 2005; Kaplan and New, 2006; Wolf et al., 2008) are likely to result in more combustible material (graminoids replaced by shrubs, deciduous mountain birch replaced by evergreen trees etc.) the frequency of fire is likely to increase still further. Burning of the insulating organic soil layer (e.g., peat) as well as the vegetation cover can lead to warming of the bare ground surface by as much as 6 °C above the mean annual air temperature, thawing of permafrost and, if ice-rich, to subsidence and ponding (Smith et al., 2008a). On the North Slope of Alaska, lightning strikes have increased tenfold since 2000 and in 2007 a fire occurred covering 1000 km^2, which was more than the sum of all known fires on the North Slope since 1950 (Jones et al., 2009; Figure 5.29). More frequent forest fires result in a change in biodiversity of vegetation (from trees to graminoids) that further affects snow trapping and albedo.

The integrated net effect of all the vegetation change feedbacks to permafrost dynamics is still unquantified, but likely to be significant.

5.3.3.3. Life in permafrost

Since the benchmark assessments of the Arctic Climate Impact Assessment (ACIA, 2005) and the Fourth Assessment of the Intergovernmental Panel on Climate Change (see Anisimov et al., 2007), there have been rapid developments in the understanding of biodiversity in permafrost and its importance.

Significant numbers (10^3 to 10^7cells per gram) of various viable ancient microorganisms are known to be present within permafrost from polar regions. In the Arctic, they have been isolated from frozen cores extracted down to 400-m depth in the Mackenzie Delta region of northern Canada (Gilichinsky, 2002). They have been extracted from permafrost in areas with high ground temperature (-2 °C) near the southern permafrost border in Siberia and also areas with low temperature (-17 °C), such as from High Arctic Ellesmere Island within the Canadian Arctic Archipelago (at 80 °N). Permafrost is a depository of viable anaerobic and aerobic, spore-forming and spore-less bacteria, green algae and cyanobacteria (Vishnivetskaya, 2008), yeast (Faizutdinova et al., 2005), and mycelium fungi (Ozerskaya et al., 2008). It also contains organic compounds resulting from metabolic activity such as the pigments chlorophyll *a*, *b* and pheophytin (Erokhina et al., 2004), biologically active intra- and extracellular enzymes (Vorobyova et al., 1997), and biogenic methane (Rivkina et al., 2007). Organs and individuals of higher systematic levels have also been found in permafrost, such as mosses (Gilichinsky et al., 2001), seeds (Yashina et al., 2002) and protozoan flagellates, infusoria and amoeba (Shatilovich et al., 2005). Although permafrost is usually regarded as an extreme habitat that limits life, its cold-adapted biomass is many times higher than that of the overlying soil cover. The age of the isolates corresponds to the duration of the permanently frozen state of the embedding strata, and date back 3 million years in northern Siberia (and probably 25 million years in Antarctica) (Gilichinsky et al., 2007). The organisms preserved in permafrost are the only life forms known to have retained viability over geological time. Current and projected permafrost thawing (Sections 5.2.2.2 and 5.2.2.3) will expose modern ecosystems and environments to relic life, with largely unknown consequences.

The activity of life forms demonstrated by microbiological and biogeochemical processes occurring below the freezing point remains open to debate. Bacteria are able to grow at temperatures below 0 °C if the medium is not frozen (Gilichinsky et al., 1993) and at least part of the permafrost community grows at temperatures between -2 and -10 °C (Bakermans et al., 2006; Steven et al., 2006). Anabolic metabolism that leads to the formation of bacterial lipids occurs down to -20 °C (Rivkina et al., 2000) and bacteria are able to carry out redox reactions between -17 and -28 °C after thousands to millions of years within permafrost (Rivkina et al., 2004). Cell viability and growth on media implies a high capacity for DNA repair in the frozen environment (Gilichinsky et al., 2008) and long-term survival is closely tied to cellular metabolic activity and DNA repair (Johnson et al., 2007). Consequently, although DNA is usually degraded rapidly in most environments, it has survived within permafrost over geological time (Willerslev et al., 2003, 2004; Lydolph et al., 2005; Hansen et al., 2006; Vishnivetskaya et al., 2006).

The occurrence of viable Cenozoic microorganisms within permafrost may provide a window into microbial life as it was before the impact of humans (Tiedje et al., 1994). If the relic and modern communities have different sensitivities to antibiotics and heavy metals, the old genes may have modern applications – or conversely, may create new problems if they are released from thawing permafrost. Bacterial and plant viruses were found in 500- to 100 000-year old ice from the Greenland Ice Sheet (Castello et al., 2005). This indicates that viruses potentially dangerous to humans could also be present in permafrost. The preservation of influenza A RNA was recently reported in ice (Shoham, 2005) and lakes on the East Siberian Sea coast that are visited by large numbers of migratory birds and that could potentially spread such viruses (Zhang et al., 2006). There is also the possibility that viruses within the bodies of people that died in epidemics and who were buried in the upper permafrost layers that are now thawing, could again become prevalent.

Conceptually, the presence of life in the permafrost could help to resolve fundamental scientific issues such as the period for which life can be preserved on Earth, or on planets such as Mars.

The overall relationship between life preserved in permafrost and current/future climate warming is complex but can be generalized as follows: only those organisms (beneficial or harmful) in the upper layers of permafrost could be released by air warming and permafrost thawing, but climate change research (e.g., ice-core drilling) and increased accessibility to areas of ancient permafrost could lead to intended or accidental access to ancient, viable life forms.

5.3.4. Feedbacks to climate through trace gas emissions and albedo changes

5.3.4.1. Albedo and evapotranspiration

Vegetation and permafrost interact in many complex ways (see Section 5.3.3). Climate warming and its impacts on permafrost will lead to significant changes in vegetation (see Section 5.3.3.1). The ability of trees and shrubs to trap snow increases albedo and short vegetation that is covered by snow in winter absorbs only about 5–15% of incident radiation (Callaghan et al., 2005; McGuire et al., 2007). In contrast, tall black spruce can intercept about 95% of incident radiation (Juday et al., 2005). Although vegetation changes to date appear to have had minimal effects on atmospheric heating in Arctic Alaska (Chapin et al., 2005), complete conversion to shrub tundra has the potential to increase summer heating by around 6 to 21 W/m^2 per decade. Complete conversion of tundra to tree cover in northern Alaska is estimated to increase summer heating by around 26 W/m^2 (Chapin et al., 2005). In addition, black carbon (soot) from more frequent wildfires (and coal burning at lower latitudes) is now falling on snow and ice, making them darker and thus less reflective (see Chapter 4, Section ???). In contrast, wildfires and ecosystems produce tiny particles (termed 'aerosols') that are carried in the atmosphere and reflect incoming solar radiation, producing a cooling effect (see Section 5.3.4.2.3). Although these feedbacks are usually measured and modeled in terms of their effects on air and/or canopy temperature, soil and permafrost temperatures are expected to follow this pattern based on geographical analogues of air/soil temperature relationships and

vegetation type. However, shading by a denser canopy may have local and temporary effects on soil temperatures.

Local disturbance and logging may also affect energy exchange with the atmosphere. For example, while fire disturbance often reduces albedo shortly after the fire, it also provides the opportunity for shrubs, and eventually deciduous broadleaf trees to develop, which will generally raise albedo. Along these lines, insect and pathogen outbreaks that result in loss of canopy foliage may cause short-term changes in albedo and start successional pathways that lead to the replacement of conifers by deciduous broadleaf trees. Thus, disturbance regimes that increase the proportion of non-forested lands and deciduous forests could reduce energy absorption and result in cooling (Chapin et al., 2000) that could to some extent preserve permafrost.

In Siberia, 55% of the coniferous forest on continuous permafrost soils is deciduous (larch, *Larix* sp.) with consequently high winter albedo. Although the two evergreen conifers (Siberian spruce, *Picea obovata*; Siberian pine, *Pinus sibirica*) also grow well on permafrost soils, *Larix* usually dominates because of its additional fire resistance. However, a recent advance of evergreen conifers into the *Larix* zone is related to increased temperature and precipitation over the past 30 years (Kharuk et al., 2005). Undergrowth of evergreen conifers decreases winter albedo, a positive feedback to climate warming and permafrost degradation. In North America, where crown fires predominate (McGuire et al., 2002), deciduous stands usually have a well-developed second canopy layer of spruce (Wirth, 2005).

After these vegetation types are disturbed by fire or logging, a sometimes sparse post-fire vegetation and the lack of a deciduous pioneer phase result in a high and sustained production of sensible heat (Schulze et al., 1999). Severe fires in the Russian Far East and probably elsewhere can cause the collapse of permafrost and prevent the recovery of trees, effectively increasing albedo by converting conifer forests into ecosystems dominated by deciduous herbs and shrubs for hundreds of years.

Changes in vegetation resulting from climate change, thawing permafrost and disturbance affect the sequestration of atmospheric CO_2 by ecosystems and evapotranspiration, both of which could result in cooling. The balance and integrated net effect of the opposing feedbacks, although likely to be significant, remain unquantified as they are difficult to model because effects vary over different scales of time and space (see feedbacks chapter). However, vegetation modeling studies to quantify impacts of climate warming in the Arctic and the role of feedbacks, suggested that forests in the eastern Canadian Arctic would have a net negative feedback on climate warming through increased sequestration of carbon, whereas forests in Arctic Russia would have a net positive feedback on climate warming through decreased albedo (Betts and Ball, 1997; Betts, 2000). More recently, vegetation models applied to the Barents region found that changes in vegetation decreased albedo by 18% (Wolf et al., 2008) but increased evapotranspiration (Göttel et al., 2008), leading to a cooling of 1 °C in spring in the eastern part of the region.

5.3.4.2. Terrestrial trace gas emissions

Exchange of carbon between the land and the atmosphere is currently mediated mainly by the generally carbon-rich active layer. However, permafrost also contains vast stores of carbon sequestered during the Holocene or even earlier in some areas. The ultimate strength of the feedback from permafrost carbon to climate change depends both on the pool size of organic carbon stored in permafrost and the rate of release to the atmosphere (Schuur et al., 2008). ACIA (2005) concluded that key issues in relation to terrestrial trace gas emissions as affected by permafrost processes include the overall carbon balance and the potential for increased CH_4 emissions. Since the Arctic Climate Impact Assessment, a new key issue has become the need to revise earlier calculations of trace gas emissions in order to correct for previously under-estimated carbon stocks. Several studies and a major review of the sensitivity of the overall Arctic carbon balance (McGuire et al., 2009) have addressed these problems. In addition, some surprising new findings have been published in relation to non-CO_2 trace gas emissions (e.g., N_2O). The following section

reports new estimates of the pool size of stored carbon, while subsequent sections address current and potential future rates of its release to the atmosphere and the possible global consequences of this release.

5.3.4.2.1. Carbon stocks

New estimates of terrestrial carbon pools are significantly higher than those presented in the Arctic Climate Impact Assessment (ACIA, 2005). Permafrost together with the active layer above it contains about twice as much carbon as is found in the global atmosphere (Schuur et al., 2008), that is, more than 1400 to 1850 Gt in the circumpolar North, including almost 300 Gt in the form of peat (McGuire et al., 2009; Tarnocai et al., 2009). Earlier studies suggested that northern peat deposits stored about 500 Gt of carbon (Gorham, 1991; Botch et al., 1995; Smith et al., 2003), with an estimated 278 Gt of this total located in permafrost regions. Most of the perennially frozen peat deposits are found in palsa and peat plateau bogs where relatively dry surfaces promote the existence of permafrost.

Recent work has shown permafrost soil carbon pools to be much larger at depth than previously recognized because of cryogenic (freeze-thaw) mixing (Bockheim, 2007; Bockheim and Hinkel, 2007) and sediment deposition (Schirrmeister et al., 2002, Zimov et al., 2006b). A detailed study of the North American Arctic quantified 98 Gt of carbon in the top 1 m of soil, an apparent 75% increase over previous comparable estimates that had been extrapolated from far fewer data (Ping et al., 2008). While the level of measurement detail represented by this North American study is not available for the permafrost region as a whole, recent syntheses of available data indicate that the entire northern circumpolar permafrost region may contain 1024 Gt of soil carbon in the surface 0 to 3 m depth, with an additional 648 Gt of carbon locked in deep layers (~25 m thick) of aeolian and alluvial 'yedoma' sediments (407 Gt), and deltaic deposits (241 Gt) of large Arctic rivers (Zimov et al., 2006a,b; Schuur et al., 2008; Tarnocai et al., 2009). This 0 to 3 m permafrost-zone soil carbon estimate of 1024 Gt represents a large fraction of global soil carbon stocks, which have been estimated to be 2300 Gt (from 0 to 3 m depth, peatlands not included) (Jobbagy and Jackson, 2000). Much of the carbon stored in Arctic permafrost regions, even old carbon (Dorrepaal et al., 2009), is highly labile and decomposes quickly under favorable moisture and temperature conditions. Although the potential release of carbon could therefore be very significant, the rate of permafrost thaw (see Section 5.2.2.2 and 5.2.3) and the subsequent changes in soil moisture are highly uncertain.

5.3.4.2.2. Current carbon and other trace gas fluxes

Many new measurements of carbon fluxes have been made since the Arctic Climate Impact Assessment (ACIA, 2005), and a complex picture emerges in which CH_4 emissions assume particular significance in terms of radiative forcing.

The annual source-sink CO_2 exchange between the terrestrial ecosystem and the atmosphere fluctuates dramatically over decadal time scales at tundra and forest sites (Oechel et al., 2000; Barr et al., 2007; Dunn et al., 2007). Spatial variability is also high across tundra sites from the European Arctic (Heikkinen et al., 2004), Siberia (Corradi et al., 2005), Alaska (Kwon et al., 2006), Greenland (Soegaard et al., 2000; Groendahl et al., 2006), Svalbard (Lloyd, 2001), and northern Scandinavia (Aurela et al., 2004; Johansson et al., 2006) and boreal forest sites (Lloyd et al., 2002; Milyukova et al., 2002; D'Arrigo et al., 2004; Wilmking et al., 2004). The variability of CO_2 exchange in Arctic tundra ecosystems is driven primarily by variability in growing-season timing and duration and moisture conditions, the latter related to permafrost occurrence and dynamics. Overall, field studies conducted to date suggest that tundra regions in the Arctic are currently sources of carbon to the atmosphere under conditions that become dry and mesic (e.g., in well-drained settings or in warm and dry years) and are carbon sinks under wet conditions (e.g., in poorly-drained settings or in cold and wet years). Regional modeling studies indicate that tundra regions across the Arctic have recently been acting as a weak sink of atmospheric CO_2 and are likely to continue to be a weak sink throughout the 21st century in response to the projected changes in climate (Callaghan et al., 2005; Euskirchen et al., 2006; Sitch et al., 2007; McGuire et al., 2009).

5.3.4.2.3. Sensitivity of ecosystem carbon balance to future permafrost thaw

H1_Carbon dioxide

Overall, the net effect of warming and permafrost thaw on CO_2 exchange is not clear: once permafrost thaws, the direction of feedbacks to the climate system depends largely on landscape wetness and dryness but this varies from region to region and with time (see Section 5.3.1). Permafrost degradation can proceed through gradual but widespread thickening of the active layer and talik formation. This will expose previously frozen deposits to anaerobic decomposition contributing to surface fluxes. In addition, landscape processes such as thermokarst erosion and fire, while more localized, can rapidly remobilize the frozen carbon pool and expose it to both aerobic and anaerobic pathways for decomposition. Hence, biogeochemical, periglacial and vegetation processes will play a key role in carbon feedbacks from ecosystems in permafrost terrain (Schuur et al., 2008).

Field and modeling studies both indicate that warming could cause release of carbon as CO_2 through enhanced decomposition in aerobic Arctic soils (i.e., soils that are not saturated with water and where the water table drops) (Oechel et al., 1995; McGuire et al., 1995, 2007; Christensen et al., 1998; Arneth et al., 2002; Dorrepaal et al., 2009) and from soils that thaw in areas of discontinuous permafrost (Goulden et al., 1998). Large releases of CO_2 are possible if landscape drying were to become pervasive in areas of continuous permafrost because ponds and wetlands comprise a substantial part of the landscape (Smol and Douglas, 2007). Indeed, a number of analyses have indicated that landscape dryness has increased over the past several decades in the boreal forest, particularly in areas of discontinuous permafrost (Smith et al., 2005b; Riordan et al., 2006; Goetz et al., 2007; Bunn et al., 2007). In contrast, CO_2 emissions from soils are likely to decrease if permafrost thaws in situations where drainage is impeded and decomposition is diminished because of anaerobic conditions (Christensen et al., 1998, 2004) and moss production is increased (Turetsky et al., 2000): permafrost thawing in Manitoba that displaces black spruce trees by *Sphagnum* moss is estimated to result in a doubling of carbon sequestration in peat (Camill et al., 2001).

The effect of thermokarst on trace gas fluxes varies over time as plant successions follow ground subsidence (Schuur et al., 2008). A study focused on upland thermokarst in Alaska demonstrated changes in plant and soil processes as a function of time since thermokarst disturbance was initiated. Increased thaw and ground surface subsidence increased net and gross primary productivity as plant growth was stimulated by thaw (Vogel et al., 2009). Species composition changed along with changes in plant growth rates, as graminoid-dominated moist acidic tundra shifted to shrub-dominated tundra with increasing thaw (Schuur et al., 2007). Increased carbon uptake by plants initially offset increased ecosystem respiration such that this thermokarst was a net sink of carbon 15 years after the initiation of thaw, even though decomposition of older carbon deep in the soil was already taking place (Schuur et al., 2009, Vogel et al., 2009). Over subsequent decades of thaw, plant growth rates remained high but increased old soil carbon losses eventually offset increased carbon uptake and this thermokarst became a net source of carbon to the atmosphere (Vogel et al., 2009). The documented emission rates suggested that 4.5 to 6.0 kg C /m^2, or 9.5% to 13% of the soil organic matter pool could be lost on a century time scale (Schuur et al., 2009). If these rates were a typical response to widespread permafrost thawing across the permafrost zone, resulting annual net carbon emissions could be similar in scale in the future to current biospheric emissions from land use change.

H1_Methane

The concept of a potential CH_4 'bomb' from wet permafrost regions has been widely recognized and there have been new insights into the dramatic dynamics of permafrost thawing in Siberia and associated expansion of thaw lakes with huge organic carbon deposits (Zimov et al., 2006a) that form local CH_4 hotspots (Walter et al., 2006, 2007a,b). Methane release is a common pathway of carbon loss in lowland thermokarst areas where drainage is restricted (Strom and Christensen, 2007; Myers-Smith et al., 2008)

and any increased release is of concern because CH_4 has a 25-fold greater heat trapping capacity than CO_2 on a century time scale (Solomon et al., 2007). Recent measurements in autumn at a High Arctic site in northeastern Greenland showed that CH_4 emissions can be concentrated in time within the seasonal cycle as well as in space: up to 50% of the total annual CH_4 emissions may take place during the freeze-up period (Mastepanov et al., 2008).

In northern Sweden, changes in permafrost dynamics, their effects on ecosystems and their feedbacks on climate in terms of increased CH_4 emissions counteract the negative feedback of increased CO_2 sequestration by vegetation (Christensen et al., 2004; Johansson et al., 2006; Figure 30). Synthesis of carbon fluxes at the catchment-scale level showed that changes in the sources of CH_4 through increased permafrost thawing may also change the sign (cooling to warming) of the current radiative forcing, due to the stronger impact per gram of CH_4 relative to CO_2. In a study of lowland thermokarst in three Canadian peatlands, carbon accumulation in surface soil organic matter was higher in unfrozen bogs and in areas where permafrost had degraded in comparison to areas where permafrost was intact (Turetsky et al., 2007), consistent with an Alaskan upland study (Section 5.3.3.1). Permafrost thaw in this lowland system promoted the release of CH_4 as waterlogged conditions predominated in *Sphagnum* spp. moss lawns that replaced the feathermoss / black spruce forest where permafrost degraded. Methane release was hypothesized to potentially offset the observed surface soil carbon accumulation for at least 70 years until plant and ecosystem succession in the moss lawn created conditions more like the unfrozen bogs that stored surface soil carbon but released smaller amounts of CH_4. These findings are consistent with the studies of decadal-scale vegetation change as a consequence of permafrost thaw in sub-Arctic Sweden (see above: Box 5.2) (Christensen et al., 2004; Malmer et al., 2005; Johansson et al., 2006).

Box 5.2. Processes in methane release from the seabed and subsequently from sea to atmosphere

The possible sources of CH_4 in the Arctic coastal seas include sediment microbial activity, natural seeps, and gas hydrate destabilization (Kvenvolden et al., 1993). Methanogenesis can occur at any depth (Koch et al., 2009). The present understanding of the mechanisms that control the current thermal state and stability of submarine permafrost and of seabed CH_4 deposits is mostly based on modeling results. These results are very controversial and suggest a wide range of possible current states of submarine permafrost (Soloviev et al., 1987; Kvenvolden et al., 1992, 1993; Kim et al., 1999; Delisle, 2000; Romanovskii and Hubberten, 2001; Romanovskii et al., 2005; Gavrilov, 2008). Saline waters affect permafrost formation (Osterkamp, 2001), and according to Osterkamp and Harrison (1985) the reduction in thickness of ice-bearing permafrost determined by the salinity of sub-permafrost waters can be hundreds of metres. In addition, Delisle (2000) predicted that open taliks can form under the warming effect of large river flows. However, Romanovskii et al. (2000) and Romanovskii and Hubberten (2001) argued against downward destabilization of subsea permafrost and suggested that upward warming through the geothermal heat flux predominated.

Destabilization becomes evident by the formation of CH_4 migration pathways through the seabed (Kvenvolden, 2002). They appear as pockmarks, mud volcanoes, funnels, chimneys, and pingo-like structures, and they might not be morphologically specified (Hovland et al., 1993; Judd, 2004; Paull et al., 2007). Additional pathways could be via submerged thaw lakes, which by the time of inundation were underlain by taliks, thereby providing a vent (Romanovskii et al., 2005). In addition, depressions found in the East Siberian Arctic Shelf bottom topography could be interpreted as a typical thermokarst terrain similar to the landscape characteristic of the Siberian Lowland (Schwenk et al., 2006; Rekant et al., 2009).

Subsea permafrost does not necessarily represent a rocklike, ice-bonded layer but is sometimes ice free under negative temperatures due to freezing-point depression by salinity, which allows gases to escape (Himenkov and Brushkov, 2007). A number of additional factors allow temporary permeability of submarine permafrost, and these include permafrost breaks due to thermal contraction, settling and adjustment of sediments, and endogenous seismicity (Osterkamp and Romanovsky, 1999). Gaseous CH_4

can escape through the seabed into the water by means of air voids, channels of unfrozen water, and fissures within the ice (Biggar et al., 1998; McCarthy et al., 2004; Arenson and Sego, 2006).

Overall, therefore, the complexity of the different radiative forcing of CO_2 and CH_4 is superimposed on the complexity of the contrasting responses of these greenhouse gases to varying effects of permafrost thaw. Hence, to assess impacts on climate, the atmospheric carbon balance must be weighed in a radiative forcing perspective (Christensen et al., 2007; Schuur et al., 2008).

H1_Biogenic volatile organic compounds

There have also been recent measurements of biogenic volatile organic compounds (BVOC), including the first report of methanol emissions (Holst et al., 2010) from sub-Arctic wetlands underlain by discontinuous permafrost (Bäckstrand et al., 2008, 2010). Biogenic volatile organic compounds can be highly reactive in the atmosphere or can form aerosols and cloud condensation nuclei that scatter and absorb radiation. Areas of permafrost thaw represented by wetland vegetation (*Eriophorum* and *Sphagnum*) showed the highest fluxes of non-methane volatile organic compounds (NMVOCs) compared with neighboring areas not undergoing thaw, represented by palsas with cold peat and plant communities with feather mosses and dwarf shrubs (Bäckstrand et al., 2008). As NMVOCs could account for about 5% of total net carbon exchange, permafrost thaw could lead to a significant increase in NMVOC emissions that affect the carbon balance of ecosystems as well as atmospheric chemistry, radiation scattering and cloud formation. The consequences of permafrost thaw via NMVOC emission are not yet quantified over large areas.

H1_Nitrous oxide

Recently, Repo et al. (2009) discovered very high emissions of the powerful greenhouse gas N_2O from 'peat circles' in permafrost regions in the Russian Arctic. Elberling et al. (2010) also reported high N_2O emissions of 34 mg N / m^2 per day in cores taken from northeastern Greenland and incubated in a laboratory, that equate to daily N_2O emissions from tropical forests on a mean annual basis. Although the importance of the process cannot be generalized across the Arctic, the strong radiative forcing potential of N_2O (298 compared with CO_2: Solomon et al., 2007) suggests important potential contributions to climate forcing that need to be calculated.

H1_Lateral transport of carbon

The vertical emission and sequestration of carbon from a particular terrestrial area is accompanied by lateral transport of carbon as dissolved organic carbon (DOC), dissolved inorganic carbon (DIC) and particulate organic carbon (POC) from land to stream to lake to river to sea. Carbon can be effluxed to the atmosphere from the water as CO_2 or CH_4, or can be exported to the ocean in the form of bicarbonates (HCO_3^-) and carbonates (CO_3^{2-}) where the carbon, initially from terrestrial areas, is sequestered over the long term (Humborg et al., 2010). Relative to other ocean basins, river transport of terrigenous carbon to the Arctic Ocean basin is about tenfold higher than average (McGuire et al., 2009). Extensive new data from previously unstudied Siberian streams and rivers over a large area of peatland in the permafrost region, suggest that mobilization of currently frozen, high-latitude soil carbon is likely over the coming century in response to the projected Arctic warming (Frey and Smith, 2005). Whereas cold, permafrost-influenced catchments release little DOC to streams, considerably higher concentrations are found in warm, permafrost-free catchments and these concentrations are related to the area of non-frozen peatland. Frey and Smith (2005) calculated that an increase in mean annual temperature to more than -2 °C would lead to 700% increases in stream DOC concentrations and 2.7 to 4.3 Tg/y (29% to 46%) increases in DOC flux to the Arctic Ocean. Already, analysis of the structural composition of DOC transported in the Kolyma River, Russia, and some of its tributaries shows a transition from loss of present-day carbon from surface horizons during spring flood to loss of old, previously stabilized carbon from deeper soil layers by the end of the summer period (Neff et al., 2006). However, in contrast, Humborg et al. (2010) suggested

that thawing of permafrost would not necessarily result in a positive feedback to warming because increased mineralization of soil organic carbon could be counter-balanced to some extent by increased weathering and DIC formation. Indeed, a decrease in DOC export relative to water discharge in the Yukon River during summer and autumn was recorded by Striegl et al. (2005, 2007) and DIC has increased (Walvoord and Striegl, 2007), thereby increasing the carbon sink term (Humborg et al., 2010).

5.3.4.2.3. Global implications of high latitude terrestrial greenhouse gas emissions

Despite recent advances in measuring CH_4 emissions, the major variations in the rate of growth of global atmospheric CH_4 cannot be explained. The atmospheric CH_4 growth rate has shown a decade of unexplained variation (down to zero) but a substantial rate of increase since 2007 (www.esrl.noaa.gov/gmd) and it is possible that part of this substantial recent increase is due to an increasing source from wetlands at high northern latitudes.

Recent studies have highlighted the importance of permafrost carbon pools in the millennium-scale development of the global terrestrial carbon cycle and its interaction with the climate system. For instance, MacDonald et al. (2006) suggested a role for rapid early development of circum-Arctic peatlands in the coincident ice core records of atmospheric CO_2 and CH_4. In addition, Walter et al. (2006) indicated that thermokarst lake formation and associated increases in CH_4 emissions contributed to post-glacial warming. Considering the future, Zhuang et al. (2006) focused on an area more extensive than the Arctic (southward to 50 °N) and projected that a current net source of carbon to the atmosphere of 276 Tg C/y would increase to 473 Tg C/y by 2100 in response to climate scenarios, permafrost thaw, CO_2 fertilization of photosynthesis, and fire. This estimate contrasts with models of the Arctic that project this area will remain, as currently, a small sink of carbon throughout the coming century (although the uncertainty is high) (Callaghan et al., 2005; Euskirchen et al., 2006; Sitch et al., 2007; McGuire et al., 2009). However, results from a coupled carbon and climate model simulation suggested that even the higher carbon source estimates from areas south of 50 °N would exert relatively small radiative forcing on the global climate system compared to anthropogenic emissions (Zhuang et al., 2006; McGuire et al., 2009).

To project the future role of permafrost carbon pools in the future global carbon cycle, a key need is to revise earlier conclusions on trace gas emissions to correct for the previously under-estimated Arctic carbon stocks (Tarnocai et al., 2009), but on the other hand, to include recent downward revisions of projected permafrost thaw (see Box 5.1). Furthermore, few of the modeling studies considered the potential effects of fire, which is a disturbance agent that has the potential for the rapid release of large stocks of carbon to the atmosphere. This release can occur not only in the direct and immediate combustion of organic matter, but also by removal of the protective organic layer above the permafrost that exposes the remaining carbon stocks to decomposition from substantial soil warming and permafrost thawing (Rocha and Shaver, 2009, in press). The fires produce black carbon (soot) that enhances snow melt and results in some degree of permafrost preservation, but the overwhelming balance of the two processes results in significant net permafrost thaw and carbon loss. In a long-term perspective, peatlands (nutrient-poor areas) in sub-Arctic permafrost regions have a lower net carbon accumulation rate than those in boreal non-permafrost regions. Hence, future warming might increase carbon sequestration in current sub-Arctic ecosystems. In contrast, ground subsidence and increased precipitation are likely to increase the proportion of fens (nutrient-rich wetlands), which historically have lower net carbon accumulation rates (Kuhry and Turunen, 2006).

Overall, it is difficult to resolve the complexity of measured variations in trace gas fluxes into projections of future radiative forcing due to permafrost thaw. Although even high estimates of radiative forcing contribute relatively little to the global system compared with anthropogenic carbon emissions (Zhuang et al., 2006; McGuire et al., 2009), recent measurements of CH_4 and N_2O fluxes suggest a future 'wildcard' (Corell et al., 2008).

5.3.4.3. Trace gas emissions from subsea permafrost

Since the Arctic Climate Impact Assessment (ACIA, 2005), there have been recent advances in measuring and estimating large-scale emissions of carbon from the Arctic coastal seas (e.g., Shakhova et al., 2010a). These new studies have resulted in much higher estimates of carbon emissions than those from earlier studies, with more profound implications for feedback to the climate system.

Arctic coastal seas host two distinct reservoirs of carbon: organic matter that could, under appropriate conditions, provide a substrate for CH_4 production and previously-formed CH_4 preserved within the seabed reservoirs some of which are permafrost. Organic-matter reservoirs consist of sediments that form during times of ocean transgressions and also of terrestrial organic matter, accumulated when the continental shelf was dry (Kleiber and Niessen, 2000; Vetrov and Romankevich, 2004: see Section 5.2.1.2). Seabed CH_4 reservoirs store natural gas, CH_4-hydrates and CH_4-bearing fluids (Soloviev et al., 1987; Ginsburg and Soloviev, 1994; Kvenvolden, 2002). Models suggest that the transgression of the Arctic Ocean over the previously exposed continental shelf during the past 10 000 years has increased the temperature of the surface sediment up to +12 °C (Soloviev et al., 1987; Kvenvolden, 1988, 1991) while measurements show that submarine permafrost is close to the thawing point in some areas (Rachold et al., 2007).

Fluxes of CO_2 and CH_4 represent significant components of the marine carbon cycle in the Arctic seas that had not previously been included in estimates of the regional carbon balance (Semiletov et al., 2007, 2011; Semiletov and Pipko, 2007; Macdonald et al., 2008; Pipko et al., 2011). Estimates of the combined CO_2 summertime evasion (for the period 2003 to 2008) from the Laptev Sea and the East Siberian Sea range between 4.2 Tg carbon (lower estimate) and 21 Tg carbon (moderate estimate) (Semiletov et al., 2010, see also Figure 3.31). Both values are of the same order of magnitude as the annual organic-matter accumulation in the Russian Arctic seas, 9 Tg C/y, a value almost equal to the gross organic-matter accumulation in the pelagic World Ocean (Vetrov and Romankevich, 2004). Decaying terrestrial organic matter dominates over marine organic matter as a source of CO_2 to the atmosphere in this area and terrestrially-derived CO_2 within the Siberian Shelf seas could be an important and so far underestimated source to the atmosphere. This source will probably increase with the projected future warming, as permafrost thawing is projected to increase, as are river discharge and coastal erosion.

Measurements of CH_4 include high sea bottom concentrations of 94 nM at the 20 m isobath in the Beaufort Sea (Kvenvolden et al., 1992) and increasing concentrations of dissolved CH_4 in the East Siberian Arctic Shelf between 1994 and 2008 that reached 900 nM in plume areas. In September 2005, about 45% of the study area exhibited concentrations of dissolved CH_4 of over 30 nM, 23% at more than 50 nM, and about 10% at more than 100 nM (Figure 5.32; Shakhova et al., 2007a,b; Shakhova and Semiletov, 2007; Shakhova et al., 2010 a,b). Cramer and Franke (2005) also measured elevated concentrations of dissolved CH_4 ($\leq$46 nM) in the middle water in the central Laptev Sea. Concentrations of dissolved CH_4 in the surface water beneath the fast ice measured in April 2007 at one location reached 5000 nM. Many bubbles were observed entrapped within the fast ice and this confirmed ebullition as the predominant mechanism of CH_4 transport through the water column (Shakhova et al., 2008a). In total, it was shown that during 2003 to 2008, more than 50% of the studied area of the East Siberian Arctic Shelf served as a source of CH_4 to the atmosphere (Shakhova et al., 2008b).

Increases in CH_4 concentration in the atmospheric layer above the sea surface reached up to 8.2 ppm in the East Siberian Arctic Shelf during 2005 (Figure 5.33), while the average mixing ratio of atmospheric CH_4 was 2.4 ppm. This measured average was comparable to average background concentrations of CH_4 measured in the atmosphere above the shallow marine sites affected by active vents and/or seepages from decaying hydrates (Shakhova et al., 2008a, 2010a,b). The increases sometimes occurred as pulses (Figure 5.33). In 2006, an increased mixing ratio of CH_4 (up to 10%) was observed up to 1800 m above the sea surface (Shakhova et al., 2008a,b, 2010a).

The destabilization of submarine permafrost has significant implications for global climate. Current estimates of the amount of CH_4 that could be released from the Arctic continental shelf (7 million km^2) during the short Arctic summer (100 days), based only on diffusive fluxes, is as high as 5 Tg of CH_4 (Shakhova et al., 2010b). This is a considerable increase on the ~0.1 Tg previously estimated by Kvenvolden et al. (1993).

The current estimate reflects the contribution of only a very small fraction of the total CH_4 fluxes and other significant components exist. One such component is CH_4 release during the deep autumn convection, which allows water from the East Siberian Arctic Shelf to mix from top to bottom (Kulakov et al., 2003). A significant late-summer potential CH_4 release to the atmosphere might therefore occur during only a few weeks (Shakhova et al., 2010b).

Another mechanism of CH_4 ventilation is deep convection in the flaw polynyas (band-like ice-free areas), which form simultaneously with land-fast ice in November. Flaw polynyas reach tens of kilometres in width and migrate out of fast ice hundreds of kilometres northward (Smolyanitsky et al., 2003), providing a pathway for CH_4 to escape to the atmosphere during the Arctic winter. Fluxes from the European Arctic polynyas are 20- to 200-fold higher than the ocean average and, as long as concentrations of dissolved CH_4 in the bottom water do not exceed 50 nM, can reach 0.02 Tg CH_4 a year (Damm et al., 2007).

A significant amount of CH_4 could also be released during the ice break-up period from areas not affected by polynyas. In these areas, dissolved CH_4 accumulates beneath the sea ice as it does in northern lakes (Semiletov, 1999). Additional release of CH_4 via these mechanisms would contribute to an increase in the diffusive fraction of air-sea CH_4 exchange, but the most important and still unmeasured component is ebullition. Assuming that ebullition might contribute to the total transport of CH_4 in the East Siberian Arctic Shelf as much as it does in northern lakes (50% to 90%), the annual release might reach from 10 to 50 Tg of CH_4. Note that this amount does not include non-gradual or sudden releases of CH_4, which are likely to take place in some areas where hydrates decay (Leifer et al., 2006).

The amount of CH_4 that could theoretically be released in the future is enormous. The volume of gas hydrates that underlie the Arctic Ocean seabed is estimated at 2000 Gt of CH_4 (Makogon et al., 2007). About 85% of the Arctic Ocean sedimentary basins occur within the continental shelf so that within the East Siberian Arctic Shelf alone, which comprises about 30% of the area of the Arctic shelf, hydrate deposits could contain around 500 Gt of CH_4. An additional two-thirds of that amount (around 300 Gt) is stored in the form of free gas (Ginsburg and Soloviev, 1994). Because most submarine permafrost is relict terrestrial permafrost, the carbon pool held can be estimated from knowledge on current terrestrial carbon storage to include not less than 500 Gt of carbon within a 25 m thick permafrost body (Zimov et al., 2006a), 2 to 65 Gt of CH_4 as hydrates (McGuire et al., 2009) together with a significant amount of non-hydrate carbon. The total amount of carbon preserved within the Arctic continental shelf is still debatable but it could be around 1300 Gt of carbon, from which 800 Gt is previously formed CH_4 ready to be suddenly released when appropriate pathways develop. Release of only 1% of this reservoir would more than triple the atmospheric mixing ratio of CH_4, probably triggering abrupt climate change, as predicted by modeling results (Archer and Buffett, 2005).

5.3.5 Socio-economic issues

Information on social and economic issues related to changes in permafrost in the Arctic is not easily available (Glomsrød, 2006). In order to assess the impacts of changes in permafrost on socio-economics, it is of paramount importance to have an understanding of the present situation, the different variables and factors that affect development scenarios. The Arctic Climate Impact Assessment examined the potential impacts of climate change on Arctic infrastructure (Instanes et al., 2005). Particular concerns were associated with permafrost warming and degradation, coastal erosion, the stability and maintenance of transport routes, and industrial development. It was concluded that adaptation, mitigation, and monitoring techniques will be necessary to minimize the potentially serious detrimental impacts.

Permafrost regions are not densely populated. Duhaime and Caron (2006) gave a broad definition of the Arctic, including land areas with discontinuous and sporadic permafrost, and estimated the Arctic population at over 9.9 million in 2002 (Table 5.1).

Table 5.1. Arctic population in 2002 in relation to permafrost area (Duhaime and Caron, 2006).

Country	Population, 1000s	Percentage of Arctic population	Percentage of country's population	Permafrost distribution
Alaska, United States	648	6.5	0.2	All types
Canada	113	1.1	0.4	All types
Greenland	56	0.6	100.0	All types
Iceland	289	2.9	100.0	Not continuous
Faroe Islands	47	0.5	100.0	None
Norway (including Svalbard)	465	4.7	10.1	All types
Sweden	509	5.1	5.7	Not continuous
Finland	645	6.5	12.4	Not continuous
Russian Federation	7144	72.1	5.0	All types
TOTAL	9915	100.0		

The country with by far the greatest proportion of the Arctic population is the Russian Federation. As a result, this country also has the majority of the infrastructure associated with population centers and the presence of permafrost (Table 5.1).

Bogoyavlenskiy and Siggner (2004) gave a more narrow definition of the Arctic, and estimated the circumpolar Arctic population at about 4 million. Of some 370 settlements in tundra regions, more than 80% are located on the coast where permafrost is commonly present. Coastal Arctic regions have concentrations of industrial facilities associated with oil and gas activities, such as the Prudhoe Bay region in northern Alaska and the Pechora Basin in Russia, as well as exploration in the Mackenzie Delta-Beaufort region with the potential for future development. There are also several mines within the permafrost regions, with exploration ongoing and the potential for further mineral resource development. In most parts of the Arctic, human settlements are dominated by relatively small communities, whereas in northern Russia there are large cities with over 100 000 inhabitants and river ports with developed urban transportation and industrial infrastructure largely serving the needs of the extracting industries.

Several towns and cities in permafrost areas of the Russian Arctic have a population of around 50 000 citizens and require a substantial infrastructure to function (Table 5.2). Further east, the Sakha (Yakutia) Republic has a population of around one million; with the capital Yakutsk being the largest city and having a population of around 200 000.

Table 5.2. Industrial cities of the Russian Arctic (population data from Russian Statistical agency 2009: www.gks.ru/bgd/regl/b09_109/Main.htm).

Region	City	Population	Main industrial activity or natural resource	Comment
Murmansk Oblast	Murmansk	311 200	Harbor, ship repair	Seasonal frost
	Severomorsk	53 500	Ship building	Seasonal frost
	Kandalaksha	36 600	Aluminum	Seasonal frost
	Apatity	61 600	Apatite	Seasonal frost
	Kirovsk	30 200	Apatite	Seasonal frost
	Monchegorsk	48 100	Nickel	Seasonal frost
	Olenegorsk	22 400	Iron	Sporadic permafrost
	Kovdor	19 100	Iron	Seasonal frost
	Zapolyarnyy	17 700	Nickel	Sporadic permafrost
	Nikel	15 000	Nickel	Sporadic permafrost

Komi Republic	Vorkuta	71 400	Coal	Continuous permafrost
	Inta	33 400	Coal	Continuous permafrost
Yamalo-Nenets Autonomous Okrug	Urengoy	118 700	Gas	Continuous permafrost
	Nadim	47 300	Gas	Discontinuous permafrost
Taymir Autonomous Okrug	Norilsk	203 900	Nickel, copper, cobalt, non-ferrous	Continuous permafrost
Sakha (Yakutia) Republic	Yakutsk	264 100		Continuous permafrost
	Neryungri	63 200	Coal	Continuous permafrost
	Aldan	23 500	Gold	Continuous permafrost
Chukotka Autonomous Okrug	Anadyr	12 600	Gold, coal, non-ferrous	Continuous permafrost
Magadan Oblast	Magadan	114 800	Gold, silver, non-ferrous	Discontinuous permafrost

Furgal and Prowse (2008) estimated that the population in the Canadian northern territories will increase from 104 000 in 2005 to 121 700 in 2031 under a moderate population growth scenario. Based on a report by the US Arctic Research Commission (2003), Alaska has a population of around 500 000 in permafrost areas, the main proportion (80%) living in sporadic permafrost areas with permafrost extent less than 50%.

The economic and strategic importance of the Arctic is high, due to the abundance of natural resources and raw materials. Military activity during the Cold War period and recent economic development has increased construction activity related to infrastructure, oil and gas facilities, transport networks, communication lines, industrial projects, and engineering maintenance systems. All of these developments have taken place with an awareness of current permafrost conditions, however, projected climate-driven changes in permafrost (see Section 5.2.3) are likely to affect both these and future developments, beyond current planning and engineering provisions.

5.3.5.1. Effect on infrastructure

Infrastructure is defined as facilities with permanent foundations or the essential elements of a community. It includes schools, hospitals, various types of building and structure, and facilities such as roads, railways, airports, harbors, power stations, and power, water, and sewage lines. Infrastructure forms the basis for regional and national economic growth.

Climate change is likely to have significant impacts on existing Arctic infrastructure and will influence the design of future development in the region (ACIA, 2005). In most cases, engineering solutions are available to address climate change impacts on permafrost, thus the issue is more economic than technological. In addition, it is possible that the uncertainty associated with projections of future climate change will increase the cost of new projects in the Arctic. Permafrost engineers must address the problem of preserving infrastructure under projected future climate conditions. One solution is to construct new buildings as existing ones are damaged and abandoned. It is possible that this method will be inadequate, since the required rate of new construction rises exponentially using the climate projections. In areas of warm, discontinuous permafrost, it is very difficult to find economic solutions to address the impacts of climate change on foundations or structures. These areas, together with the coastal zone containing saline permafrost, where the combined problems of increased wave action, sea-level rise, and thermal erosion have no simple engineering solutions, present the greatest challenges in a changing climate.

5.3.5.1.1. Buildings

Engineering problems associated with warming and thawing of permafrost are well documented. They include deformation and damage to buildings and development of thermokarst in northern Russian cities. In June 2002, an apartment building was ruined in the small town of Chersky in the lower Kolyma Region. It happened due to thawing permafrost under the foundation. In summer 2006, at the parking place in Yakutsk, several cars fell into a huge thermokarst crater caused by thawing permafrost (André and Anisimov, 2009). However, these impacts may be associated with warming of the ground that can

accompany construction and operation of infrastructure and inappropriate engineering design rather than the direct effects of a changing climate.

It is often not possible to differentiate effects on constructions damaged, between those effects related to increased permafrost temperature due to changes in climate, and those resulting from changes to the ground caused by other factors (Instanes, 2003). For example: site conditions are often different to the assumed design site conditions; design of the structure did not take into account appropriate load conditions, active-layer thickness and permafrost temperature; the contractor did not carry out the construction according to the design; the maintenance program was not carried out according to plan; and, the structure is not used according to design assumptions.

Although there has been limited scientific evidence to date that observed climate warming has been the direct cause of failure of engineering structures on permafrost, new studies are suggesting that this effect is sometimes important. A study of historical air temperature in Spitsbergen, central Yakutia, and Alaska (Instanes and Anisimov, 2008) suggested that piled foundations have not suffered sufficient loss in bearing capacity to become unstable due to the observed increases in air temperature. However, the loss in capacity of piles in the period 2000 to 2007 could be the early warning of potentially detrimental impacts of a changing climate. In addition, an increase in the active layer thickness may induce frost-jacking of installed piles and this may present a bigger threat to pile and foundation stability than loss of bearing capacity.

Oberman (2003a) performed a case study for multi-storey buildings in the city of Vorkuta in northwestern Russia. Buildings constructed on thawed soil (taliks) show very little deterioration with time (line 5 and 6, Figure 5.34). For buildings constructed on permafrost (lines 1 to 4) the deformation coefficient progressively increased with time. Oberman (2003a) reports that one 5-story residential building was abandoned less than 10 years after it was constructed and will be demolished, although the expected lifetime of the construction was 50 years (Figure 34).

Catastrophic deformations of buildings are confined to the 1980s. These are the years characterized by the greatest temperature increase in permafrost composed of Quaternary mineral deposits and peat. According to Romanovsky et al. (2010a) this increase in permafrost temperature is due to the combination of increasing air temperature and increasing snow depth (Figures 5.35 and 5.36).

Six to seven years after construction, the deformation coefficient of buildings is increasing rapidly. This is an indication that engineering-geocryological conditions are changing rapidly and are not the same as assumed in the initial design considerations. According to Oberman (2003a), the reason for such changes in conditions can only be natural degradation of permafrost.

In addition to effects of thawing permafrost on more-or-less flat areas of continuous permafrost, projected increases in temperature, precipitation, and storm magnitude and frequency are very likely to increase the frequency of avalanches and landslides (see Section 5.3.2), while slow downslope movements over time can also have implications for infrastructure. In some areas, the probability of severe impacts on settlements, roads, and railways from these events is very likely to increase. In some coastal areas, rockfalls might lead to local tsunamis thereby widening the impact on coastal structures. Structures located on sites prone to slope failure are very likely to be more exposed to slide activity as groundwater amounts and pore water pressures increase. An increasing probability of slides coupled with increasing traffic and population concentrations is very likely to require expensive mitigation measures to maintain a defined risk level. The best way to address these problems is to incorporate the potential for increasing risk in the planning process for new settlements and transport routes.

5.3.5.1.2. Pipelines

The conclusion from a recent study of the permafrost-related performance of the Trans-Alaska Pipeline since its inception in 1977 was that global climate change is not expected to threaten its future operation or integrity (Johnson and Hegdal, 2008). However, increased maintenance may be expected. Similar conclusions have been drawn from Canadian studies. Currently, there are three small-diameter pipelines operating in northern Canada, with the longest being the 869 km oil pipeline transecting the discontinuous permafrost zone from Norman Wells in the Northwest Territories to Zama, Alberta. The Norman Wells pipeline, in operation since 1985, is an ambient line and is the only oil pipeline in North America that is completely buried in permafrost. An extensive monitoring program carried out by the pipeline operator and the Canadian Government (e.g., Naviq Consulting Inc and AMEC Earth and Environmental, 2007; Smith et al., 2008a; Burgess et al., 2010) concluded that although climate change was not considered in the design, thaw penetration and thaw settlement beneath the right-of-way has remained generally within design values. Climate change effects have been largely obscured by the effects of vegetation clearing and thermal effects of the pipe (Burgess and Smith, 2003; Smith et al., 2008b), especially during the first decade of operation as permafrost was responding to an abrupt change in ground surface temperature of about 2 °C. Warming of permafrost is occurring in the adjacent undisturbed terrain which is consistent with rising air temperatures (Smith et al., 2005a) and these effects of ongoing climate warming on the pipeline right-of-way are projected to become more apparent over the longer-term, becoming the dominant influence after 25 to 50 years of pipeline operation (Smith and Riseborough, 2010).

In contrast to the situation in Alaska and Canada, the rapid expansion of pipeline construction in Russian permafrost areas has caused problems with above-ground pipelines due to thawing of ice-rich soils and frost heave of pipeline foundations (Perlshtein, 2008). On average, about 35 000 failures are registered annually affecting the 350 000 km long network of pipelines in western Siberia: more than 20% are most probably due to deformations and weakening of foundations induced by permafrost thaw (Anisimov and Reneva, 2006). Uplift of 1.5 m due to frost heave has been observed in a single year along the Urengoy pipeline (Pazinyak, 2001). These effects are likely to be related to construction (i.e., right-of-way clearing and leveling, etc.) and to operation of the pipelines and not directly linked to climate change.

Many of the failures in Arctic Russia occurred on marginal pipelines connecting the extracting facilities at specific locations with larger hubs partly because construction and maintenance of smaller pipelines are under relatively loose control compared to main transportation lines. Nevertheless, damage to such pipelines may have dramatic environmental impacts. In 1994, the break of the pipeline connecting the *Vozei* oilfield extracting facilities with the hub 'Golovnyue Souruzhenija', resulted in a spill of 160 000 tons of oil-containing liquid, the world's largest terrestrial oil spill. Soil contamination in the surrounding area is still high, 15 years later. Although a Russian commission concluded that the cause of the leak was corrosion, surveyors and local operating engineers argued that it was caused by differential thaw settlement which was estimated to be 0.5 to 1.5 m (Oberman, 2007). Similarly, thermokarst and uneven ground settlement led to significant deformation of a 45 km-long test pipeline section in the Pechora region after a few years of construction. Although it is debatable whether the real cause was inappropriate engineering design, the experience illustrates high vulnerability of constructions built in sporadic permafrost environments to changes in the thermal state of the frozen ground (Oberman, 2003b).

5.3.5.1.3. Roads, railroads and runways

Changes in weather and climate extremes can have considerable impact on transportation systems (Peterson et al., 2008). In permafrost regions, the possible climate impacts on road operations are related to degradation of the pavement structure due to permafrost warming and thawing. This can lead to reduced traffic speed, higher accident risk (landslide), and road closure.

Kondratiev (2008) presented an evaluation of recent construction of railroads and highways on permafrost in Russia and China over the past 10 years. For more than 100 years, construction experience of roads and railroads on ice-rich permafrost soils shows that thaw settlement and frost heave have occurred with or without climate warming. The lifetime of a road on ice-rich permafrost is relatively short, typically less

than 20 years. It is therefore expected that road systems will be upgraded regularly and that a gradual adaptation to climate change is possible. In Alaska, the runway serving the Prudhoe Bay oilfields has been reconstructed due to settling from thawing permafrost (Hinzman et al., 2005). In Nunavik, permafrost degradation is threatening the integrity of roads and airfields (Ministére des Transports, 2005).

Since its completion in 1975, the Svalbard airport runway has experienced pavement unevenness mainly caused by thaw subsidence (and consequent frost heave) of the ice-rich soil layers in the embankment (Instanes and Mjureke, 2005). Observed climate warming was not found to be responsible for the recent damage to the runway and the projected climate scenarios do not pose an immediate threat. However, compared to the current situation, maintenance costs will probably increase during climate warming, but it is possible to gradually adjust the runway to a warmer climate by applying insulation or more frequent maintenance.

In the polar regions, the stable solid platform provided by frozen ground is utilized for winter transportation, and also for resource exploration and construction activities. Thawing of permafrost and changes in the length of the thaw season could have an effect on economic activity in these regions through changes in the seasonal scheduling and shortening of the period over which these activities may take place (Prowse et al., 2009). The timing of the freezing and thawing of the active layer will be important, as will frost penetration in non-permafrost areas within the discontinuous permafrost zone. It is anticipated that the length of the freezing season will decrease in response to climate warming and during extreme, warm years that the thaw season may be extended (e.g., Atkinson et al., 2006; Smith et al., 2009a).

5.3.5.1.4. Effluent storage

In a changing climate, structures need to maintain their integrity over periods of many decades to centuries as structural failure may have significant consequences, such as the release of contaminated effluents into the surrounding environment, with consequent impacts on ecosystems and human health. Remedial action may be required to maintain the structural integrity of such sites. For high-risk projects such as water retaining structures, tailings and storage facilities for hazardous waste, the environmental protection agencies can apply the principle of 'perpetual' design (discussed further in Section 5.4).

5.3.5.2. Natural resources

5.3.5.2.1. Fossil fuels

The Arctic contains large amount of fossil fuels and is an important supplier of oil and gas to the global market (Table 5.3).

Table 5.3. Arctic petroleum reserves as a percentage of the global total (Lindholt, 2006).

Reserves	Oil	Gas
Production	10.5 %	25.5 %
Proven reserves	5.3 %	21.7 %
Undiscovered resources	20.5 %	27.6 %

According to Lindholt (2006), 97% of total Arctic oil and gas production is located in Russia and Alaska (Alaska contributes 20% of total US production). Most fields are located onshore, but there are substantial reserves in the Beaufort Sea and Pechora Sea. In addition, Canada, Greenland and the Barents Sea are potential locations for future major oil and gas activities. The undiscovered resources of global oil and gas reserves are estimated at 20.5% and 27.6%, respectively (Lindholt, 2006). The United States Geological Survey has estimated the undiscovered resources of oil and oil-equivalent natural gas at 412 000 million barrels of oil equivalent (Bird et al., 2008).

The Arctic Monitoring and Assessment Programme has published an assessment of oil and gas activities in the Arctic (AMAP, 2007). The main findings relevant to the present study are that oil and gas activity in the Arctic is likely to increase and that the technology and use of best practices have lowered environmental impacts to date, but that additional risks may occur as conditions change or new areas are explored and developed. It is possible that climate change will have both positive and negative financial impacts on the exploration, production, and transportation activities of this industry.

Oil and gas are mainly transported within the Arctic through pipelines and in tankers. Climate change impacts on oil and gas development have so far been minor, but are likely to result in financial costs and benefits in the future. For example, the warming and thawing of permafrost, on which buildings, pipelines, airfields, and coastal installations supporting oil development are located, is very likely to adversely affect these structures and the costs of maintaining them (see Section 5.3.5.1).

5.3.5.2.3. Mining

The Arctic holds large stores of minerals, ranging from coal to gemstones and fertilizers (Lindholt, 2006) and is an important contributor of raw materials to national and global economies. It is also expected that the Arctic contains large reserves of undiscovered resources of raw materials. However, many known reserves are not exploited at present due to inaccessibility.

Russia extracts the largest quantities of minerals, including nickel, copper, platinum, apatite, tin, diamonds, and gold, mostly on the Kola Peninsula but also in Siberia (see Table 5.2). Three diamond mines are currently operating in the Canadian Arctic, and it is likely that mining activity there will increase in the future (Prowse et al., 2009). The main deposits include coal, cobalt, copper, barium, beryllium, bismuth, diamonds, gold, iron, lead, lithium, nickel, niobium, silver, tantalum, tungsten, uranium, and zinc. Coal and gold mining continues in Alaska, along with the extraction of lead and zinc deposits from the Red Dog Mine, which contains two-thirds of the US zinc resources.

Mining activities in the Arctic are likely to benefit from improved transportation conditions, although the actual extraction process is unlikely to be much affected by climate change. However, climate change effects on permafrost will possibly affect those maintenance and waste and tailing containment facilities that rely on frozen conditions to isolate contaminants from the environment. Mining facilities with roads on permafrost are likely to experience higher maintenance costs as the permafrost thaws (Instanes et al., 2005). Any expansion in oil and gas activities and mining in the Arctic is likely to require the expansion of air, sea, and land transportation systems (see Section 5.3.5.1.3).

5.3.5.3. Relocation of communities

Inuit communities are located in predominately low-lying coastal zones, and many already have to take action to protect shorelines and buildings, and to consider future relocation as a result of encroaching erosion and existing damage (Furgal and Prowse, 2008). Four indigenous communities in Alaska are planning relocation due to coastal erosion and flooding (Bronen, 2008). However, according to Bronen (2008) this process is problematic for various reasons, including: the lack of a government agency with the authority to relocate communities; lack of funding specifically designed for relocation; lack of criteria for selecting relocation sites; and the lack of a governmental organization responsible for the strategic planning of relocation. As a consequence, while adaptation is technically possible, governance issues are hindering appropriate responses to climate change impacts on coastal permafrost.

5.4. Strategies for adapting to climate change effects in permafrost regions

- For more than 100 years, construction experience of roads and railroads on ice-rich permafrost soils shows that thaw settlement and frost heave have occurred with or without climate warming.

- The added problem of climate-driven permafrost thaw now requires more maintenance of existing structures, and more effective design and practice with a forward vision for new structures.

- There is an increased acknowledgment and consideration of climate change and its impacts on permafrost in engineering design, particularly for structures for which there are high consequences of failure. New maps based on a probabilistic approach of infrastructure susceptibility to permafrost thaw in Eurasia show that large areas of Russia, especially in a zone along the coast have a high susceptibility of buildings and engineered structures to ongoing changes in climate and permafrost.

- Much of the literature on geohazards, social science and policy in polar regions, fails to adequately address issues related to the effects of climate change on permafrost, so developing adaptation strategies could be difficult.

Thawing permafrost is just one of many stressors that will lead to environmental change to which we need to adapt. Direct adaptations to thawing permafrost *per se* involve infrastructure, whereas other impacts of thawing permafrost, such as impacts on ecosystems, require adaptations (to opportunities and problems) that also involve globalization, changes in culture, and socio-economic issues. This assessment addresses direct adaptations of infrastructure development that arise from thawing permafrost induced by climate change. The impact that Arctic communities can have on mitigating future changes in climate, and therefore permafrost, is likely to be insignificant. As mitigation at a global scale is likely to be a slow process, Arctic communities, governance and organizations outside the Arctic but with involvement in Arctic infrastructure, must therefore adapt. Some of these adaptations will be to processes that take place over long periods and that are continuous over time, whereas rapid changes (e.g., following threshold exceedence and extreme events) may even require the development of emergency contingency plans.

Construction work and infrastructure development in ice-rich permafrost is very likely to cause thawing of the underlying ice-rich permafrost and frost heave problems if proper engineering solutions are not applied. The Fourth Assessment of the Intergovernmental Panel on Climate Change acknowledged this problem with the statement "*Although several recent scientific and media reports have linked widespread damage to infrastructure with climate change [...], the effect of heated buildings on underlying ice-rich permafrost can easily be mistaken for a climate-change impact*" (Anisimov et al., 2007). For example, the majority (but not all: see Sections 5.3.5.1.1 and 5.3.5.1.2) of damage to structures in the Russian permafrost regions in the period 1980 to 2000 resulted mainly from poor maintenance rather than climatic change (Kronik, 2001). The added problem of climate-driven permafrost thaw now requires more effective design and practice with a forward vision. However, much of the literature treating geohazards, social science and policy in polar regions fails to adequately address issues related to the effects of climate change on permafrost (Nelson et al., 2002), so developing adaptation strategies could be difficult.

Furgal and Prowse (2008) stated that adaptation of northern infrastructure to climate change will mainly involve approaches already in use to reduce thermal and mechanical impacts of ground disturbance, see for example Instanes (2003), Instanes and Instanes (2008), and Kondratiev (2008). Adaptive responses require monitoring in order to evaluate infrastructure performance, to determine if changes in permafrost condition deviate from those predicted, and to decide whether additional adaptation measures are required.

The US Arctic Research Commission Permafrost Task Force (2003) listed the following strategies related to infrastructure in permafrost areas and global warming:

- To consider climate change, as predicted by global circulation models, weighed by associated probabilities, to make decisions regarding new infrastructure on permafrost more credible.

- To fund a denser network of environmental monitoring.

- To substantially increase federal funding for contaminant research in cold (permafrost) regions.

In the construction industry, various methods have been suggested to address temperature-related foundation problems. Techniques to reduce warming and thawing, such as heat pumps, convection embankments, thermosyphons, winter-ventilated ducts, and passive cooling systems, are already common practice in North America, Scandinavia, and Russia (Instanes et al., 2005). For example, a majority of the 80 flat loop thermosyphon foundations examined in northern Canada have demonstrated that this design provides an economical foundation system that will function for some time even with climate warming affecting the permafrost (Holubec et al., 2008). Instanes and Instanes (2008) advocated a foundation system in permafrost regions using a heat pump system. The system has been used on six different buildings on continuous permafrost in Longyearbyen and Sveagruva, Svalbard. It was found to be a good solution for buildings with on-ground (no elevation of the structure above ground) foundations over ice-rich permafrost soils. These types of foundation require insulation material underneath the concrete floor slab and a cooling system underneath the insulation to prevent thawing of the underlying ice-rich permafrost. This type of foundation design may also be beneficial to mitigate the effect of possible future climate warming on the structure. The heat pump cooling system is designed to decrease the temperature of the soils supporting the permafrost foundations to a design value. The heat extracted from the ground is used for heating the building above.

Facilities constructed several years ago and which need to maintain their integrity over long periods present a particular challenge as they were not designed for the warmer conditions of today or for those projected for the future. Remedial action may therefore be required if additional thawing of permafrost associated with climate warming, and subsequent changes in soil strength and settlement exceed the original design values. This may include modifications to pad or embankment design or the use of passive cooling to ensure frozen conditions in the future (Prowse et al., 2009). Waste-containment facilities that depend on encapsulation of waste in permafrost, ice-cored water retention structures, or dams to isolate contaminants from the environment are examples of structures that may require remedial action to maintain their structural integrity. The effects of climate change have been incorporated in the design of such structures in Canada since the mid-1990s. For example, structures to contain mine tailings or related to reclamation of DEW line sites (e.g., Hayley, 2004; Hayley and Horne, 2008; Prowse et al., 2009). Designing for climate change can include the use of techniques to promote and maintain frozen conditions (such as passive cooling devices like thermosyphons) and improved cover designs to provide additional insulation or the use of convective cooling, as was used for waste rock piles at the Diavik diamond mine in the Northwest Territories, Canada (e.g., MEND, 2004, Arenson et al., 2007; Pham et al., 2008a,b). Other measures can include employment of a geomembrane layer that will also provide defense against loss of containment (Hayley and Horne, 2008).

5.4.1. Facilitating adaptation at the regional scale

Comprehensive high-resolution permafrost scenarios are needed to develop strategies of adaptation to the effects of warming on infrastructure in the Arctic on a regional scale. Many of the geomorphological processes resulting from thawing permafrost are relatively well studied and may be predicted using process-oriented models coupled with scenarios of climate change. Smith and Burgess (2004) mapped the sensitivity of permafrost in Canada to warming and characterized areas in terms of the potential for thaw settlement and therefore highlighted areas where impacts of warming may have implications for infrastructure design. A numeric index was suggested by Nelson et al. (2002) to evaluate the potential threats to engineered structures due to warming and thawing of permafrost. The more recent study by Anisimov and Lavrov (2004) used a modified hazard index that also includes soil salinity, which is particularly important in the vicinity of the Arctic shoreline.

The basic assumption behind these 'settlement indices' is that the potential threats to infrastructure appear when seasonal thawing propagates deeper into the ground, and that high ice content and salinity of the soil increase the susceptibility of the existing structures to such threats. A recent map of the distribution of the hazard index over northern Eurasia, calculated using the 2050 GFDL-based climatic projection under the

IPCC B2 emissions scenario, is illustrated in Figure 5.37 (Anisimov and Lavrov, 2004; Instanes and Anisimov, 2008).

Hazard index calculations partition the permafrost region into areas with 'low', 'moderate', and 'high' susceptibility of the infrastructure to climate-induced change. A zone in the high-susceptibility category extends discontinuously around the Arctic Ocean, indicating high potential for coastal erosion. Large parts of central Siberia, particularly the Sakha Republic (Yakutia), and the Russian Far East show moderate or high susceptibility.

Particular concern is associated with the Yamal Peninsula, which falls into the highest risk zone because of the ongoing expansion of the oil and gas extracting and transportation industry into this region. Despite the relatively low temperatures, frozen ground in this area is already unstable, largely because of its high salinity, and thus even small increases in temperature may cause extensive thawing of permafrost and ground settlement. Areas of lower susceptibility are associated with mountainous terrain, landscapes in which bedrock is at or near the surface, and permafrost with low ice content.

Hazard index maps such as the one shown in Figure 5.37 provide support in decision-making with regard to strategies of adaptation to projected changes, including the various engineering applications discussed previously. Particular solutions depend on the regional permafrost projection, are structure-specific, and are often based on cost-benefit analysis. Life expectancy should be incorporated in the design of each construction, as should aging of the structures as this amplifies the effect of a decrease in the load-bearing capacity of the frozen ground.

Until recently, the intrinsically deterministic nature of models that neglect the stochastic variability of permafrost parameters was one of the factors complicating realistic evaluation of potential threats to the infrastructure at regional geographical scales. Unlike these conventional models, the new type of models that have recently been developed (Anisimov et al., 2002; Anisimov, 2009) take into account the probabilistic nature of climatic projections and small-scale spatial variability of permafrost parameters, such as ground temperature and active-layer thickness. Aside from portraying the level of uncertainty on maps representing spatial distribution of permafrost parameters, output from a stochastic model can be used to construct a series of maps depicting the probability of the parameters to exceed given thresholds within specified regions.

Probabilistic maps deliver important information supplementing traditional maps, which depict only 'average' or 'typical' values. The probability approach requires the construction of several maps for a region, each for a given threshold. The more thresholds used, the better the set of probabilistic maps in representing the spatially distributed probability density function (PDF) of permafrost parameters.

An example of a probabilistic active-layer thickness map for the northern Eurasian permafrost region under the current climatic conditions is given in Figure 5.38 (Anisimov, 2009). Similar maps may be constructed for permafrost temperature, which is a key parameter that is accounted for in the design of pile foundations in permafrost regions. Such maps have direct implications in predictive permafrost hazard assessment. They indicate the proximity of the current state of permafrost in any region to the threshold beyond which structures designed for prescribed climatic conditions may become unstable.

5.4.2. Adaptation at the local scale

Esch and Osterkamp (1990) summarized engineering concerns related to permafrost warming (Table 5.4). The Arctic Climate Impact Assessment addressed these challenges and typical engineering projects that are likely to be affected by climate change (Instanes et al., 2005).

Table 5.4. Climate change impact and related engineering problem.

Climate change	Engineering problem	Adaptation suggestion

impact		
Thawing of permafrost (development of taliks)	Decrease in effective pile length in permafrost Progressive mass movements (landslides) Progressive surface settlements	Increase pile length Relocation Artificial cooling of foundations
Increased active-layer thickness	Thaw settlement during seasonal thawing Increased frost heave forces on piles Increased total and differential frost heave during winter	Insulation Increase pile length, protection / treatment of pile in active layer to reduce frost heave forces
Warming of permafrost at depth	Increased creep rate of existing piles and footings Reduced ad-freeze bond strength for piles Increased creep of embankment footings	Increase pile length, and depth and area of footings. Reduce foundation load. Artificial cooling of foundations.

5.4.2.1. Permafrost engineering design criteria for climate change

In engineering science, there are generally two main design approaches: the deterministic approach and the probabilistic approach.

In the *deterministic approach*, fixed values are assigned to input parameters and calculations are performed with one set of values to arrive at a solution; sensitivity analyses may be performed by changing the fixed values and comparing solutions. In permafrost regions, most of the smaller and more basic projects are being analyzed using deterministic methods, but work is advancing on defining appropriate parameters.

The *probabilistic approach* recognizes that uncertainty exists in the solution to a problem. The uncertainty can be attributed both to the deterministic model and to the inherent variability of the input parameters that support the calculation. In permafrost regions, larger projects with more agency oversight are moving toward more probabilistic methods and in some cases risk-based analysis. Meyer (2008) recently advocated a risk-oriented, probabilistic design procedure in order to define design characteristics for components that could be affected by changing environmental and climatic factors.

The design lifetime of a structure in permafrost areas typically varies between 15 and 75 years (Table 5.5). Buildings and infrastructure constructed on permafrost and designed for permafrost conditions are, therefore, unlikely to be adversely impacted during typical engineering design lifetimes. However, the concept and requirement of 'perpetual design' is becoming more common. For high-risk projects such as water-retaining structures, tailings and storage facilities for hazardous waste, the environmental protection agencies can apply such principles to the design.

Table 5.5. Typical design lifetime for structures in permafrost areas.

Structure	Lifetime
Roads	15 to 20 years
Oil and gas pipelines	30 years
Buildings	30 to 50 years
Railroads	50 years
Bridges and underpasses	75 to 100 years

5.4.2.1.1. Risk-based evaluation

Bush et al. (1998) presented a methodology for considering the impact of projected climate change within the framework of the engineering design process; see also Hayley and Horne (2008). They also explained how the same methodology can be used to identify and prioritize concerns about existing facilities with respect to climate change impacts. The method involves a multi-step approach that first assesses the sensitivity of a given project to climate change and then the consequences of any potential failures (Bush et al., 1998). The relationship between sensitivity and consequences defines the risk that climate change

poses to the project. The degree of sensitivity and the severity of the consequences are used to determine what level of climate-change impact analysis should be carried out for a given project (Table 5.6).

The sensitivity of a particular infrastructure project to climate change is determined by a number of factors, including initial soil and permafrost temperature, the temperature dependence of the material properties, the project lifetime, and the existing over-design or safety margin that might be included in the design for other reasons.

Recently, Hayley and Horne (2008) described a screening process for engineering projects that includes a procedure for qualitative assessment of climate sensitivity, based on soil type and permafrost temperature. This procedure has been applied to engineering projects in northern Canada. Currently, this procedure does not have official status in Canada, but the method provides a useful tool to evaluate the climate sensitivity of engineering projects.

Table 5.6. Engineering risk levels (Bush et al. 1998).

Likelihood (Bush et al., 1998)	Frequent	Probable	Occasional	Remote	Improbable
Probability of occurrence in one single year (Instanes, 2003)	50%	20%	10%	1%	0.01%
Predicted number of occurrences (Instanes, 2003)	Once in 2 years	Once in 5 years	Once in 10 years	Once in 100 years	Once in 10 000 years
Consequence					
Negligible	C	C	D	D	D
Minor	A	B	B	C	C
Major	A	A	B	B	C
Catastrophic	A	A	A	A	B

Risk level A: Detailed quantitative analyses are required. Refine input parameters with additional investigation and testing. Perform full-scale monitoring program with periodic evaluation of performance. Independent expert review required.
Risk level B: Semi-quantitative (or limited quantitative) analyses are required. In addition to requirements for C, perform limited quantitative analysis. Use engineering judgment for input parameters. Monitor permafrost performance. Perform full quantitative evaluation for projects with limited precedence.
Risk level C: Qualitative analyses are required. Apply expert judgment. Document result of evaluation. Perform quantitative evaluation for projects with limited precedence in design, function, or construction method.
Risk level D: Analyses not required. No action required.

This approach requires site-specific analyses. Several recent site-specific analyses of the effect of climate warming on engineering projects are available in the literature (Clarke et al., 2008; Hayley and Horne, 2008; Instanes and Anisimov, 2008; Nishimura et al., 2009a,b; Zhou et al., 2009). These analyses give an indication of how specific structures respond to climate warming so that their climate sensitivities can then be determined.

5.4.2.1.2. Probability of occurrence based on output from general circulation models

An alternative approach to developing permafrost engineering design criteria for climate change is to use the output from downscaled general circulation models (GCMs) to construct scenarios of future air temperature at given locations between the present time and a set time in the future, such as 2100. These data can be used to investigate how the probability of occurrence of active-layer thickness and permafrost temperature at depth changes with time and climate scenario (Instanes, 2003). In this manner, climate warming can be treated like any other environmental load (such as earthquakes, wind, waves, currents). It is believed that this approach could also incorporate the concept of 'perpetual design'.

5.4.2.1.3. Monitoring

Monitoring of existing structures on permafrost, follow-up of historical data, and back-calculation of known failures are important to gain a better understanding of long-term behavior of structures on

permafrost. For existing projects such as the Norman Wells pipeline in northern Canada, monitoring programs are important not only for facilitating the mitigation of environmental impacts associated with the project (Naviq Consulting Inc and AMEC Earth and Environmental, 2007; Prowse et al., 2009; Burgess et al., 2010) but also for mitigating the effects of ongoing climate change. The monitoring results and their dissemination in publicly available databases (Smith et al., 2004, 2008c) are also important for the design of future hydrocarbon projects, including the incorporation of climate change. Recognition of the importance of baseline information on permafrost condition and ongoing monitoring of permafrost thermal state, led the Canadian Government to enhance the permafrost monitoring program in the Mackenzie region. This ensured that impacts related to hydrocarbon resource development in the region, including those associated with climate change, would be minimized. These efforts have already generated key data (Smith et al., 2008d, 2009b) that are being incorporated in project design and contributing to the regulatory process.

5.4.2.1.4. Heritage

Permafrost preserves ancient life forms. It also preserves more recent records of human culture and biota that are of considerable value to archaeologists and paleontologists, respectively. As permafrost thaws, these unique records will be destroyed. An adaptive response is to survey areas of thawing permafrost from the perspective of interrogating and, if possible, conserving its archives of past environments, cultures and biota: this is currently being done to document the numerous tombs of the 2500 year-old lost Scythian civilization that are preserved by permafrost in the Altai mountains (Goossens et al., 2007).

5.5. Uncertainties and recommendations

5.5.1. Data needs and priorities for permafrost research

- Permafrost dynamics are observed in many ways, both in the field and remotely. However, there is insufficient interaction among the various observation communities. *There is a need to better integrate observing techniques including the further development of remote sensing to complement* in situ *observations and expansion of* in situ *observations to validate and explain information derived from remote sensing and modeling.*

- Field based observation sites and networks are sparse relative to the geographic extent and environmental complexity of the Arctic. However, these are extremely important as they provide the data (permafrost temperature, ice content, active-layer thickness etc.) to understand baseline conditions and to characterize change. They are also essential to reduce uncertainty in predictions of future conditions and field data are also used to better understand how permafrost conditions influence the biophysical environment and processes. *We urgently recommend the maintenance and enhancement of field observation sites and networks.*

- There is yet no comprehensive circumpolar assessment of thermokarst extent. While thermokarst development has been quantified in many small areas, confidence in the rates of change is limited by the sparse time series of historical imagery and the sensitivity of the measurements to seasonal phenology and interannual differences in precipitation. There is a lack of sufficient data and models to predict the distribution of ground ice, a dominant factor controlling the magnitude of thaw settlement. In addition, there are no satisfactory methods to remotely sense permafrost degradation in ice-poor soils. *Further progress in assessing the response of thermokarst terrain to climate change requires large advances in data acquisition, monitoring, and modeling as well as integration of permafrost, geomorphic, hydrological, thermal, ground-ice, and ecosystem-succession models to address the full range of complex biophysical interactions.*

- There are considerable uncertainties in modeling future permafrost distribution and dynamics. These include an under-representation of the ice content and the organic layer and its importance in

insulating permafrost during climatic warming. Permafrost models also fail to adequately represent the disequilibrium that has arisen because some current permafrost is related to past climates. This results in a lag period between a climatic change and a response of the permafrost. *We recommend therefore, that priority be given to acquiring baseline information of specific relevance to model refinement.*

- There is too little integration of climate-cryosphere research. *We recommend the development of integrated research programs that include other components of the cryosphere, such as hydrological systems with permafrost and snow permafrost interaction, to give a holistic understanding of the climate-cryosphere system.*

- Permafrost landform dynamics and their climatic and local controls are poorly understood relative to permafrost temperatures and active-layer thickness. *It is necessary therefore, to extend permafrost observations to include permafrost landform dynamics and their drivers.*

- The underlying sources of important oscillations in atmospheric methane concentrations are poorly understood, but are most likely to be related to northern wetland dynamics. *It is therefore necessary and important to gather more data for year-round* in situ *emission dynamics.*

- Arctic carbon stocks have previously been underestimated. *A key need is to revise these earlier estimates and to project the future role of permafrost carbon pools in the global carbon cycle.*

- Measurements of carbon emissions from permafrost areas on land and sea are biased to vertical fluxes. However, there is also important lateral transport of carbon, such as in polynyas. Seasonal observations of such transport are hindered by sea-ice formation and movement. *Techniques need to be developed and deployed to measure lateral fluxes of carbon in the shelf environment.*

- Changing permafrost conditions are likely to have considerable socio-economic effects but it is difficult to bridge the gap between the science of permafrost and the assessments of socio-economic impacts, which are studied by different communities with different cultures of publication. *There should be a strengthening and integration of the projections of changing permafrost with socio-economic aspects of thawing.*

5.5.2. External use of improved understanding of permafrost

- Although there is a current move to refine the inclusion of knowledge of permafrost processes in climate models, *there is a need to set research priorities for developing the use of permafrost-related products in larger-scale climate and earth system models in a coupled manner.*

- While permafrost-related data are generally available to the scientific community, particularly within the discipline, other science disciplines and local stakeholders are often unaware of data on permafrost changes, their implications, and their uncertainties. *We recommend that the availability of permafrost-related data be increased for use by a variety of users, particularly those planning adaptation, including those outside the research community, such as engineers and land-use planners.*

- Research into impacts of thawing permafrost often needs to work at spatial and temporal scales that are currently inadequately represented in permafrost research. For example, responses of ecosystems and biogeochemical cycling as well as infrastructure may respond to temporal thresholds and extreme events while development of adaptation strategies is spatially a local process that requires local permafrost projections. *We recommend that more priority be given to downscaling models of permafrost change to scales appropriate to local decision-makers and that projection of impacts of changing permafrost focus more on responses to thresholds and extreme events.*

5.6. Conclusions

5.6.1. Opportunities and challenges

There is a growing awareness of the increased thawing of permafrost and its importance (ACIA, 2005; Solomon et al., 2007). However, many uncertainties existed in the projections of permafrost condition, its responses to climate warming, and its impacts particularly on greenhouse gas emissions. For example, there was a realization of the potential importance of carbon fluxes from degrading terrestrial permafrost and particularly the 'wildcard' of potential methane release from subsea permafrost, but this was poorly quantified. In addition, the observing networks of permafrost thermal state and active-layer thickness were sparse and integration of results was tentative.

Since the benchmark studies of the Arctic Climate Impact Assessment (ACIA, 2005) and the Fourth Assessment of the Intergovernmental Panel on Climate Change (Asimov et al., 2007), there have been many advances in the understanding of permafrost responses to climate warming, although some uncertainties remain. Some of these advances have appeared through the intensive program of permafrost research during the International Polar Year (IPY) and the Ninth International Conference on Permafrost (for which the proceedings have been published, see Kane and Hinkel, 2008). In particular, a coordinated international effort during the IPY has led to a more complete and updated baseline of the current permafrost thermal state and a characterization of changes that have occurred over the past 20 to 30 decades throughout the Arctic.

Major advances include recent observations of increased methane levels in the atmosphere above the sea surface in the East Siberian Sea, reaching 8.2 ppm during 2005, and estimates that more than 50% of the East Siberian Arctic Shelf area studied served as a source of methane to the atmosphere between 2003 and 2008. Furthermore, there has been a significant upward revision in estimates of carbon stocks occurring in terrestrial permafrost. Both advances have significant implications for the potential amplification of climate warming. In addition, models projecting change in permafrost condition have been significantly improved by including deeper permafrost as well as the organic layer, resulting in lower estimates of future permafrost thaw. Interactions between climate-driven thawing of permafrost and biological processes are complex but nevertheless understanding has improved. In particular, there is a new recognition of the importance of permafrost in preserving past life, sometimes in a viable form.

5.6.2. The human face of permafrost change

This assessment explores the socio-economic implications of thawing permafrost together with the possibilities for adapting to its impacts. Major new messages are that the technology is available to adapt buildings and other infrastructure to thawing permafrost, and that previously recorded incidents of infrastructure failure were probably more related to engineering and design problems than to climate change. There has also been an increased effort in recent years to incorporate climate change into the design of major structures.

An important concept that needs emphasizing is that the science of permafrost needs to progress simultaneously at two different spatial scales. First, the large regional scale is important to integrate the impacts of permafrost thawing on feedbacks to regional and local climate: this information, because of the potential size of the climate forcing and its global impacts should be used to drive global mitigation responses. Second, Arctic residents and those organizations outside the Arctic but with economic responsibilities/opportunities within the Arctic, need to adapt to local changes in permafrost and their consequences. As this adaptation is a local process, local projections are increasingly required from the permafrost research community.

5.6.3. 'Winners and losers'

In contrast to other components of the changing Arctic cryosphere such as snow (see Chapter 4), thawing permafrost offers few opportunities but many challenges. Again, unlike the impacts of changing snow cover, thawing permafrost will result in few 'winners'. However, when permafrost is completely lost at the southern margins of the current permafrost zone, communities will benefit in that they will no longer require special engineering design of infrastructure related to permafrost, and activities and land uses operating south of the current permafrost boundary (e.g., forestry and agriculture) are likely to migrate northward in some areas. Another possible advantage is that an increase in winter base flow in rivers could alleviate the need for freshwater supplies to large towns in Arctic Russia during winter. Overall, most of those involved in the Arctic can be regarded as 'losers': the large multinational industries and developers will face greater economic investment to stabilize infrastructure over longer periods, whereas individual residents will face disruption to communication routes and even resettlement in some cases.

5.7. Acknowledgments

The authors are grateful to the Swedish Department of Environment, the Swedish Environmental Protection Agency (grant number F-159-08) and the Swedish Research Council FORMAS (grant number 204-2009-45) for funding the involvement of the convening lead authors and for facilitating the involvement of Russian contributors. We also thank other national (United States, Canada, Norway) agencies for funding the participation of their respective lead authors. Many authors contributed written text, figures, tables and ideas and they are explicitly acknowledged on the title page, but we thank them here also. The AMAP SWIPA Secretariat gave valuable support throughout, and we thank the SWIPA Integration Team as well as authors of other SWIPA chapters for their pertinent discussions and encouragement. Finally we thank the numerous reviewers.

Appendix 5.1. Sources of information and data

Observational data

Observations of permafrost condition, including its thermal state and active-layer thickness are essential baseline data requirements for the design of Arctic infrastructure, to improve models to predict future permafrost condition, and to improve assessments of climate change impacts. Although there has been some development in the use of remote sensing of permafrost and related phenomena (Zhang et al., 2004; Duguay et al., 2005), most permafrost-related data, especially historical data, are from *in situ* measurements. These data meet a variety of user needs but their point-based nature and non-uniform spatial distribution presents some limitations, including utilization for spatial modeling purposes.

A key provider of this type of information is the Global Terrestrial Network for Permafrost (GTN-P) which was established under the Global Climate Observing System. GTN-P consists of two components (Burgess et al., 2000): the Thermal State of Permafrost (TSP), which conducts measurements of ground temperature in boreholes; and the Circumpolar Active Layer Monitoring (CALM) network, which provides measurements of active-layer thickness (e.g., Shiklomanov et al., 2008).

The GTN-P provides ongoing data on permafrost condition and the summary data are posted periodically on the websites for GTN-P (www.gtnp.org) and CALM (www.udel.edu/Geography/calm). The monitoring network was enhanced during the International Polar Year (Figure 5A.1) through the establishment of new permafrost observatories to address geographical gaps (Romanovsky et al., 2010b). An important dataset is the IPY snapshot which provides a database (Romanovsky et al. 2010b) and map (see Figure 5.8) that summarizes the thermal state of permafrost for the IPY period, 2007 to 2009 and provides a baseline against which future change can be measured in the Arctic. Data products associated with the IPY are also being developed by various national organizations (see for example, Smith et al., 2008b; Midttømme et al., 2008).

The CALM network is a global network of sites at which data on active-layer thickness and dynamics are collected (Brown et al., 2000; Nelson et al., 2008). CALM was established in the early 1990s to observe and detect the long-term response of the active layer and near-surface permafrost to changes in climate. The CALM network incorporates 168 sites in Arctic, sub-Arctic, Antarctic, and mountainous regions. Several sites constitute longitudinal and latitudinal transects across northwestern North America, Europe and the Nordic region, and northeastern and northwestern Russia. About 70% of sites are located in Arctic and sub-Arctic lowlands in the zone of continuous permafrost. Regions of discontinuous and mountainous permafrost contain 20% and 11% of sites, respectively (Shiklomanov et al., 2008).

Historical permafrost data often exist within research data collections held by diverse institutions, including academia, government agencies and national and international data centers. The International Permafrost Association (IPA) has endeavored to compile much of this disaggregated data through the Global Geocryological Database (GGD); a metadata collection describing data in diverse locations around the world (Parsons et al., 2008). Building on from the GGD, the World Data Center for Glaciology at the National Snow and Ice Data Center in Boulder, has published two compendia of permafrost related data: Circumpolar Active-layer Permafrost System (CAPS) version 1 (IPA DIWG, 1998) and version 2 (IPA SCDIC, 2003). The IPY snapshot database is an update of these earlier data products. The data products provide information on monitoring site parameters, soil temperature, cyrosols, and climatologies, as well as maps, metadata and bibliographies. They provide reasonable data snapshots and some time-series data but they are not current as the GTN-P databases will be when fully functional. Considerable efforts exist in various countries through government agencies to provide national/regional, freely accessible and downloadable information related to permafrost and of interest to a variety of users (see for example, Midttømme et al., 2008; Smith et al., 2004, 2009b).

There is likely to be a wealth of historical information that could be compiled into digital databases, but while some of these datasets will have been well managed, the utility of others may be limited (Parsons et al., 2008). Part of the reason for this is that the datasets may have been largely in the possession of individual researchers that are no longer available to provide documentation regarding site characteristics, measurement methods, instrumentation, accuracy and precision of measurements. Periglacial process data are also increasingly being collected but there is a lack of coordinated data management arrangements. However, some progress is being made in the management of process data, for example by the Norwegian IPY project and the associated NORPERM database (Midttømme et al., 2008).

Other limitations of existing data sets, such as those for the active layer and permafrost thermal state, also include a lack of continuity in the data record or a short data record. In addition, there is an uneven distribution of study sites and a number of regional gaps exist. Although various analytical and modeling techniques are available to fill both temporal and spatial gaps (through for example spatially distributed models, Sazonova and Romanovsky 2003; Sazonova et al. 2004; Romanovsky et al., 2007), these synthetic datasets are not generally available.

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6. Changing Lake and River Ice Regimes: Trends, Effects and Implications

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Key Findings

- Paleolimnological records of some High Arctic lakes indicate that recent (past 100 year) reductions in ice conditions are significant and associated with profound changes in biological productivity and diversity.

- The end of the past century has seen a decline in systematic observations of ice conditions in the Arctic and changes in observational techniques, making circumpolar assessments of ice phenology difficult.

- While ice-covered locations in more southern regions have been shown to be the most thermally sensitive to warming, recent evidence indicates that changes in ice cover duration have been greater at some High Arctic locations than in regions farther south.

- Reductions in lake-ice duration will modify thermal conditions that can lead to (i) enhanced evaporation and potential loss of shallow lakes and (ii) enhanced mixing, which can turn these Arctic systems into sinks for contaminants.

- Loss of ice cover will also lead to increased methane emissions from Arctic lakes and levels of exposure to ultraviolet radiation that exceed those due to atmospheric ozone depletion, although these could be mitigated by catchment processes producing changes in water budgets and organic inputs.

- Changes in ice-cover dynamics will alter the frequency and magnitude of extreme ice-jam floods, thereby affecting riparian and under-ice ecology as well as hazards to Arctic communities and infrastructure.

- Decreases and loss of river- and lake-ice transportation routes are producing large costs for Arctic communities and industry, and these are likely to increase in the future. Moreover, economically viable adaptation options are extremely limited.

Summary

Unlike the other major global cryospheric components that either blanket large expanses (e.g., snow, permafrost, and sea ice) or are concentrated in large volumes in specific locations (e.g., glaciers and ice sheets), lake and river ice are interwoven into the terrestrial landscape through the major freshwater flow and storage networks. As such, freshwater ice has enormous significance to physical, ecological and socio-economic systems. In discussing its role, it is important to note that the hydrological influence on this cryospheric component extends well outside the Arctic via the main northward-flowing rivers.

Surface-based observations have provided much of the information about the response of lake and river ice to climate. The earliest long-term ice observations on Arctic lakes and rivers date back to the 17th and

18th centuries. Since the middle of the 20th century, national programs of ice-thickness measurements have been carried out in most Northern countries. Unfortunately, there appears to have been a significant decline in a number of ice-observation programs over the past two decades. Moreover, there tend to be fewer observations carried out at higher latitudes as opposed to more southerly locations. To supplement current observations and to extend the historical record, a number of other methods have been used, including isotopes, tree scars, and buried taxa. Unfortunately, a transition to remote sensing by regular observational programs has not yet been realized in any circumpolar country.

Given that ice-covered freshwater bodies of various sizes represent a considerable fraction of the northern high-latitude sub-Arctic and tundra landscape, they need to be considered in any form of climate modeling of Arctic terrestrial environments. The degree to which the atmosphere is influenced depends on several factors, including the magnitude, timing, location, and duration of ice cover. A significant shortening of the freshwater ice duration period can have profound impacts on local, regional, and even larger-scale climate over the Arctic. The formation, growth, decay and break-up processes of ice cover differ markedly between lakes and rivers, but all are influenced by similar climatic variables that control surface heat fluxes. In addition to changes in such variables, other indirect controls, such as landscape hydrology, can significantly affect freshwater-ice phenologies. Hence, through modifications of terrestrial hydrology, climate also plays an indirect role in affecting freshwater-ice phenologies.

The longest-term records of freshwater ice changes in the Arctic have been derived from paleolimnological analyses. Remote Arctic lakes have the greatest potential as natural monitors and recorders of past variation in ice cover. Over past millennia, the Arctic has experienced appreciable change to lake and river networks brought about by changes in other cryospheric components, including wide-scale glacier recessions and thawing permafrost. Paleo-historical analyses of lakes from selected sources around the circumpolar Arctic indicate that proxies of lake primary production remained low throughout the middle and late Holocene, in agreement with cold prevailing conditions and prolonged ice cover. Pronounced increases in whole-lake productivity, however, occurred during the warm period of the Holocene Thermal Maximum, consistent with decreased ice cover and its influences on in-lake dynamics. Ice-cover duration then steadily increased in concert with Neoglacial cooling but was interrupted during the Medieval Warm Period (MWP), when it became equally short or shorter than today. Longer periods of ice cover returned during the Little Ice Age (LIA) when Arctic lakes became less productive than during the MWP – the dominant control on productivity being the duration of the ice-free season. Paleolimnological evidence from some Arctic lakes suggests that longer ice-free seasons have been experienced at various times since the beginning of the 19th century. Additional evidence of accelerated lake ice melt comes from lakes that had been perennially ice-covered or characterized by extended periods of ice cover. It has been inferred from some records that many Arctic lakes may have crossed an important ecological threshold as a result of recent warming.

Based on observational records of freshwater ice, numerous trend analyses have been completed, mainly focused on relatively simple characteristics such as the timing of autumn freeze-up and spring break-up, ice-cover duration, and ice thickness. A compilation of long-term records across the Northern Hemisphere for the period 1846 to 1995 indicates that freeze-up has become later by an average of 6.3 days per 100 years and break-up has become earlier by an average of 5.8 days per 100 years, resulting in an average reduction in ice duration of 12.1 days per 100 years. Only one site, however, was located north of the Arctic Circle, reflecting the lack of high-latitude, long-term observation sites. Subsequent analysis of a smaller set of lakes around the Northern Hemisphere (again, non-Arctic) with records extending to 2004/05 indicates that the average rate of change in both events was noted to increase from +6.3 to +10.7 days per 100 years for freeze-up and -5.8 to -8.8 days per 100 years for break-up, thereby further reducing average ice duration by 12.1 to 19.5 days per 100 years.

Many shorter-term regional studies have also been conducted, but they exhibit appreciable spatial and interdecadal variability. Most trends toward shorter freshwater-ice duration over much of the circumpolar North closely correspond to increasing air temperature trends and timing of the 0 °C isotherm. Broad

spatial patterns in ice trends have also been linked to major atmospheric circulation patterns, different phases of which can cause contrasting ice conditions across individual continents and between opposite sides of the circumpolar North.

Some important south-north contrasts have also been identified in freshwater-ice trends. Examples from Finland and Sweden show greater sensitivity to warming at the more temperate latitudes although contrasting latitudinal results have been noted for south-north regions of Canada. The degree to which this reflects the effects of either more recent or higher latitude warming or a combination of both is unclear. Comprehensive, large-scale records of river- and lake-ice thickness are rare. One dataset compiled for Canada over the past 50 years does not reveal any obvious trends over the latter part of the 20th century. However, when comparing the past two decades to the previous 30-year period, reductions in maximum ice thickness have been observed on nearly all rivers and lakes within Arctic Russia. Overall, explaining the reasons for any of these regional contrasts requires further investigation into regional climatic differences and potential incompatibilites of different observational methods.

Most degree-day projections of future freshwater-ice regimes indicate a continued reduction in freshwater-ice duration in all northern regions. However, empirical relationships between freshwater-ice dates and air-temperature indices may not be reliable under future climatic conditions. Few future projections have been made using global or regional climate models, one analysis of lakes between 40° and 75° N for the period 2040 to 2079 projects an expansion of the summer stratification period: 5 to 20 days earlier freeze-up, 10 to 30 days earlier break-up, and an overall decrease in lake-ice duration of about 15 to 50 days. Maximum lake-ice thickness was also projected to decrease by 10 to 50 cm. Owing to the projected increase in winter snowfall, high-latitude regions are likely to develop increased surface white ice, especially as the ice cover also thins.

Few attempts have been made to project future changes in river-ice cover; however, projected decreases in south-north gradients in air temperature suggest that the severity of break-up and related ice-jam flooding may be reduced on some large, northward-flowing rivers. This would have major implications for northern infrastructure and riparian ecosystems. Changes in the magnitude of the spring snowmelt, however, will mitigate the effect of changing air-temperature gradients on break-up conditions. Location and severity of river-ice break-up could also be modified by changes in hydraulic gradients, particularly in coastal river deltas affected by rising sea level. With continued winter warming, dynamic ice jams caused by mid-winter break-ups will increasingly intrude into higher latitudes; there is already evidence of such events occurring in the sub-Arctic.

Decreases in lake-ice duration combined with higher temperatures during the increasingly long open-water period will lead to increased evaporation and lowering of lake levels. Some very shallow northern basins are likely to dry out and possibly become athalassic systems (inland saline lakes), unless there is a compensating increase in precipitation. Recent evidence suggests that this may have already happened to some ponds that have been permanent water bodies for millennia. By contrast, decreases in ice-cover thickness are likely to increase the unfrozen water volume and available under-ice habitat, particularly in those lakes that currently freeze to the bed. Where such thinning ice receives enhanced snow-loading from projected winter increases in precipitation, pulsing of lake water into downstream river channels will increase.

Aquatic ecology of lakes and rivers will be affected by an array of interacting processes resulting from changes to freshwater-ice regimes. In general, adjustments in the timing of freeze-up and break-up on lakes will affect a wide range of related biological aspects of seasonality. Some changes are likely to be gradual, others abrupt as systems pass critical ecological thresholds. Changes are also likely both to have positive and negative effects. Of particular concern is the change in river dynamics associated with ice break-up, which has been shown to be critical to the ecological health of river deltas.

Two socio-economic sectors that will be affected by changes in high-latitude freshwater-ice regimes are transportation and hydroelectric energy production. Changes in ice regimes will reduce the capacity for transport on ice roads that are critical to the resupply of remote communities and mining centers, which cannot use air access for the transport of heavy loads. For Northerners with a traditional subsistence-based lifestyle, ice-based travel may become increasingly hazardous and reduce the ability to undertake some traditional harvesting methods. As the ice cover thins and the season is reduced, initial adaptation measures for the continued use of major ice roads could include modifications to ice-road construction techniques or transport schedules. However, continued warming will preclude ice roads as a major component of northern transport, and will necessitate the development of alternative forms of transportation, such as land-based road or rail networks, the capital costs for which are likely to be significant. Hydroelectric operations will both benefit and be challenged by changes in ice conditions that affect the rivers in which they operate and provide regulation. Monitoring and mitigating such problems will be a particular issue for hydroelectric power producers with remote facilities. The importance of river ice on hydroelectric operations may also be indirectly affected by future energy adaptations.

To facilitate comparisons of river- and lake-ice observations from around the circumpolar latitudes, there is a need for an international circumpolar effort to assemble and compile a comprehensive archive of existing information and the adoption of standardized methods of *in situ* observation. Moreover, a special focus should be placed on adopting remote sensing approaches to augment the *in situ* observation networks. This would require some form of international collaborative effort, one that could perhaps be undertaken by an international agency such as the World Meteorological Organization. Once the archive is established, a time series of spatial trends in ice phenologies should be conducted and links made to controlling trends and patterns in climate.

Fully assessing how climate has and will affect river-ice processes can only be accomplished by developing more advanced ice break-up models that consider future combined changes to landscape hydrology, in-stream hydraulics, and ice mechanics. While advancements are being made in modeling changes in lake ice, more validation efforts are required across a range of hydro-climatic regimes and lake sizes, and modeling needs to be expanded to include related effects on lentic (still water) and lotic (moving water) ecosystems. Model validation will also need to be undertaken, especially to help identify key non-linear and/or step changes.

Direct and indirect effects of changes in freshwater ice regimes will both have cascading impacts on socio-economic systems. The suitability of adaptation options can only be properly assessed through a broad range of cost-benefit analyses and additional socio-economic modeling, which also take into account cultural values held by Northern residents. Given the importance of many ice-affected socio-economic sectors, key locations of such activities should be considered in the selection of long-term observing and validation sites for lake and river ice. This is likely to maximize the socio-economic benefits of conducting future freshwater ice and climate change research in the Arctic.

6.1. Introduction

- Lake and river ice are interwoven into the terrestrial landscape through the major freshwater flow and storage networks and as a result have enormous significance for the physical, ecological and socio-economic systems within the Arctic.

- Freshwater ice plays an influential role in all terrestrial water storage and flow networks in the Arctic, although the hydrological influence on northern lake- and river-ice regimes originates well outside the Arctic, such as via the headwaters of the large northward-flowing rivers.

- River ice is the source of major hydrological extremes on rivers in cold regions, producing annual peak water levels that frequently exceed those under open-water conditions even though at much lower discharge.

- The Arctic contains a variety of types of ice-covered lake, the genesis and evolution of which are largely dependent on other components of the cryosphere.

- Northern lakes tend to be most abundant in glaciated, permafrost peatlands where they can occupy a significant part of the terrestrial landscape.

6.1.1. Objectives, methodology and structure

The overall objective of this chapter is to evaluate the effects of climate on lake- and river-ice regimes in the Arctic, as well as the effects and implications of changes in these regimes for the related physical, ecological and socio-economic systems. For this report, the geographical boundary of the Arctic has been liberally defined to ensure that material from more broadly defined circumpolar northern environments (which have high relevance to Arctic conditions) is included. Having a broader geographical focus is particularly important for this chapter because many of the ice-covered flow systems entering the Arctic originate at latitudes well to the south of the Arctic Circle.

This report generally focuses on literature published since, or not covered by, the Arctic Climate Impact Assessment (ACIA, 2005). Although some aspects of lake and river ice were addressed by ACIA (e.g., Walsh et al., 2005; Wrona et al., 2005) and in assessments by the Intergovernmental Panel on Climate Change (IPCC) (e.g., Fitzharris et al., 1995; Anisimov et al., 2001, 2007), there has never been a comprehensive international treatment focused primarily on these two cryospheric components. The present review attempts to fill this gap and at the same time advance understanding of how lake- and river-ice regimes have changed and are projected to change under altered climatic regimes. Given the more substantial treatment of lake and river ice in the present report, in addition to the new post-ACIA material included here, a substantial amount of literature that predates the ACIA report on subjects that were previously missed or only given cursory attention is also included. Such subjects cover not only the physical characteristics of lake and river ice and their association with climate, but also a suite of related effects, particularly those with ecological and socio-economic implications. In addition, and where possible, information about potential adaptation options is also included.

The remainder of Section 6.1 contains background material to place in context the importance of change in freshwater ice regimes in the Arctic. This includes a brief review of the physical nature (Section 6.1.2), geographical extent (Section 6.1.3), and socio-economic importance (Section 6.1.4) of lake and river ice in the circumpolar North. Subsequent sections of the chapter address: the sources of information about lake and river ice; the role of lake and river ice in the climate system; how lake- and river-ice regimes have changed historically and are projected to change in the future; the major future effects of changes in lake- and river-ice regimes on the physical, ecological, and socio-economic systems of the Arctic as well as potential adaptation options; and the major uncertainties and need for future research.

6.1.2. Physical description of lake and river ice

Ice-covered freshwater ecosystems are dominant features of the terrestrial environment of the near-polar latitudes. This section briefly reviews the characteristics of freeze-up, growth, and break-up of lake and river ice at high latitudes. For more details see the reviews by Adams (1981), Ashton (1986), Beltaos (1995a, 2008a), and Prowse (2005), as well as later sections of this chapter.

Where open-water conditions prevail during the summer months, an ice cover will first begin to form along the margins of lakes and rivers as the water column cools from net energy losses to the atmosphere. For lakes and rivers, the efficiency of autumn cooling depends on surface area to volume ratios. Thus, large shallow lakes and rivers cool more rapidly than their small deep counterparts. Under calm conditions on lakes, a stable ice cover will eventually extend over the entire water surface and, with additional heat loss, begin to grow vertically downward, forming relatively transparent (typically referred to as 'black',

‘blue’, or ‘congelation’) ice. The rate of ice growth per unit of heat loss will decrease as the ice thickens and progressively insulates the underlying water column from the atmosphere.

In contrast to lakes, ice forms under much more dynamic conditions on rivers. Although large sheets of ice may develop on large, slow-moving rivers, the more common progression to freeze-up involves the formation of frazil ice, which begins as microscopic ice particles within the turbulent flow and evolves to various surface forms (such as frazil slush and pans) as the ice continues to grow and accrete. Transported downstream, these floating pans will eventually reach a sufficiently high surface concentration that they will bridge across and constrict a river cross-section (Figure 6.1a). Other incoming ice will then accumulate behind such ice and cause freeze-up to advance upstream. Depending on flow velocities and the strength of the accumulating ice, the initial ice cover may collapse, shove downstream, and thicken until it has sufficient internal strength to resist the application of forces from upstream. As in the case of lakes, subsequent ice growth will then largely occur by downward vertical growth.

Growth of lake and river ice continues throughout the winter as heat is progressively lost from the water column. Depending on the severity of the winter and the depth of the water body, some lakes and rivers within the Arctic will totally freeze to the bed, thereby acting as a barrier to water flow and related aquatic biota. By contrast, some Arctic rivers can have sections that remain open all winter largely due to strong turbulence or local heat sources, such as from groundwater. These are the freshwater equivalent of marine ‘polynya’.

Ice growth may also be influenced by surface snow accumulation, which can act as an additional surface insulator and hence slow ice growth. It can, however, also promote the growth of surface ice through the formation of ‘snow’, ‘white’, or ‘slush’ ice. This occurs when sufficient snow accumulates on the ice surface to depress the existing ice cover below its hydrostatic water level, thereby flooding or ‘slushing’ some of the surface snow. Subsequent refreezing of this slushed snow leads to white-ice formation, which can be rapid compared to the deeper growth of further black ice, because it is more directly exposed to atmospheric cooling. White ice tends not to be a major feature of Arctic conditions and is more prevalent at temperate latitudes because snow loads are, for the most part, relatively small. The exception is where snow accumulates early in the season when the ice cover is relatively thin. White ice (as noted in Section 6.3.2.1.), is an important feature of lake- and river-ice covers because of its strong effect on ice-surface and water-column radiation regimes.

Icings (also known in German as ‘aufeis’ for ‘ice on top’ or ‘naled’ in Russian) are another form of freshwater ice that forms in river channels and which is similar in structure to the previously described white ice. In general, icings are surficial accumulations formed by water flowing onto the surface of an ice cover or ground, usually as a result of some localized flow resistance in a channel that forces water to the surface or, in the case of smaller rivers and streams, from groundwater seepage (e.g., van Everdingen, 1990). They can accumulate as a widespread sheet of ice or as a localized mound depending on the rate of flow versus surface freezing. In cases where the process continues throughout the winter and where an icing deposit may be fed by a number of seepage points, entire channels and even floodplains can be coated to thicknesses several times the depth of the open-water channel or the thickness of ice that would occur from normal freezing of the underlying flow. Some icings in Russia are estimated to be hundreds of kilometres long, 0.5 to 1.0 m thick, and more than 1000 m wide (e.g., Sokolov, 1986; Sokolov et al., 1987). In terms of icings formed by groundwater seepage, the total volume contained in Russian Arctic river basins alone is estimated at 100 km^3 (Sokolov, 1986). Large icing accumulations can pose a flood hazard, especially where they plug narrow flow channels, such as through culverts, or where their mass decreases the flow until it overtops banks (Prowse, 2005). In a somewhat similar fashion, large quantities of frazil ice may accumulate to form a potentially dangerous type of ice jam, the ‘hanging dam’ that can also lead to enhanced flooding in the spring (Ashton, 1986). Typically, however, most spring flooding occurs as the result of break-up of floating ice covers.

As with freeze-up, the break-up of freshwater ice tends to be more dramatic on rivers than on lakes (Figure 6.1). Both ice covers go through a pre-breakup period of ablation: first losing surface layers of snow, then ice-surface melting, and in rivers, accelerated thinning of the ice bottom as flow velocities increase. At very high latitudes where the ice ablation season is very short compared to the season for ice growth, freshwater ice on lakes may exist in a perennial form, developing only moats of summer meltwater. Freshwater ice can experience significant reductions in mechanical strength brought about by the internal absorption of solar radiation, particularly during the spring as incoming radiation seasonally increases and the highly reflective snow layers are ablated (e.g., Hicks et al., 2008). This can be important to the dynamics of break-up on rivers, which typically generate the most important hydrological event of the year (see Section 6.5.1.1). Water levels during spring break-up often far exceed those possible under open-water conditions at equivalent levels of discharge, as shown in Figure 6.2a for a station on a major tributary of the Mackenzie River (de Rham et al., 2008a). For the entire Mackenzie basin, almost half (13 of 28) of the hydrometric stations were found to have annual peak water-level events occurring exclusively under ice break-up conditions. As illustrated in Figure 6.2b, latitude was not found to be a major controlling factor, and other physical influences, such as elevation and slope, seem important. However, even for sites dominated by open water, ice conditions were found to be important because they can significantly elevate water levels during the spring break-up period. For example, as illustrated in a dimensionless stage versus discharge plot of all station types (Figure 6.2c), a spring break-up flow (Q_I) of only 10% of the open-water flow (Q_O) produces a nominal water depth of at least 50% of that for open-water conditions. Within the totally ice-dominated regime of the curve, a flow equivalent to only 25% of the open-water discharge will produce an equivalent nominal water depth during spring break-up. At the extreme end, an equivalent open-water discharge will produce an approximate 50% increase in nominal water depth during the spring break-up event. Such a curve exemplifies the importance of river ice in the generation of extreme high-stage events on northern rivers. As detailed in later sections, such events can be detrimental to the built environment but are also crucial to the ecosystem health of many Arctic aquatic environments, particularly riparian zones and river deltas.

6.1.3. Spatial extent

Unlike the other major global cryospheric components that either blanket large expanses (e.g., snow, permafrost, and sea ice) or are concentrated in large volumes in specific locations (e.g., glaciers and ice sheets), lake and river ice are interwoven into the terrestrial landscape through the major freshwater flow and storage networks. As such, freshwater ice has enormous significance to a variety of physical, ecological and socio-economic systems. In discussing the areal extent and volume of Arctic freshwater ice, it is important to note that the hydrological influence of this cryospheric component extends well outside the Arctic. For example, the major northward-flowing rivers of North America (Mackenzie) and Russia (Lena, Ob, Yenisey) have their headwaters in more temperate latitudes, and ice-related effects in such locations have the potential to influence downstream Arctic environments. In the case of river ice, Bennett and Prowse (2010) estimated that about 56% of the northern-hemisphere river network, extending southward to 33° N in North America and 27° N in Eurasia, experiences conditions conducive to some ice formation (Figure 6.3). Moreover, cold conditions existing for half the year are found on river headwaters as far south as 28° and 27° N in these two continents, respectively. For large rivers in cold continental regions, such as the Lena and lower Mackenzie, or at high latitudes, such as the Yukon, ice conditions can persist for more than six months over the entire river length. By contrast, for rivers with more temperate headwaters, only sections (e.g., 73% of Ob River length) experience such long-term ice effects.

The spatial extents of temporal ice regimes in lakes are expected to be similar to those described for rivers. A large proportion of the above-noted regime classifications, however, are found outside the polar latitudes and, for the most part, lakes within these southern regions do not play a direct role in the Arctic. The exception would be lakes belonging to the hydrological networks draining to the Arctic, most notably including those such as Great Slave Lake, Lake Athabasca, Lake Baikal and Great Bear Lake, although the latter is partly found north of the Arctic Circle.

The Arctic contains a variety of lake types, their genesis and evolution largely dependent on components of the cryosphere, particularly glaciers, permafrost and river ice. Examples include post-glacial lakes remaining from the Pleistocene or that have evolved in the deglaciated environment, thermokarst lakes and ponds, and rarer forms, including karst, meteoritic impact crater, stamukhi (Section 6.5.2.3), tectonic and volcanic lakes (e.g., McNight et al., 2008; Pienitz et al., 2008). In the case of areas north of around 45° N, northern lakes tend to be most abundant in glaciated, permafrost peatlands (with 14.4 lakes per 1000 km^2) and least abundant in unglaciated, permafrost-free terrain (with 1.2 lakes per 1000 km^2) (Smith et al., 2007). Overall, thermokarst lakes and ponds represent the most abundant and productive aquatic ecosystems in the Arctic (Vincent et al., 2008b). They are most common in flat-lying regions underlain by fine-grained, ice-rich sediment (French, 2007), and are found extensively in the lowland regions of western and northern Alaska (Hinkel et al., 2005), Canada (Cote and Burn, 2002; Marsh et al., 2009) and Siberia, where they comprise about 90% of lakes in the permafrost zone (Walter et al., 2006). See Section 6.3.1 for further discussion about lake coverage in relation to climatic influences.

6.1.4. Physical, ecological and socio-economic importance

The Arctic Climatic Impact Assessment established freshwater ice cover as an important component of Arctic hydrology that influences many physical, chemical, and biological processes operating in lentic and lotic systems (Walsh et al., 2005; Wrona et al., 2005). Lake ice directly affects many limnological properties and processes, including solar radiation inputs and their spectral signature for photobiological and photochemical processes; ultraviolet radiation; air-water gas exchange; water-column heat budgets; stratification and under-ice mixing; biogeochemical dynamics; and the entrainment of terrestrial inputs, including contaminants (e.g., see review by Vincent et al., 2008b). River ice exerts a similar broad range of controls on lotic systems, including the productivity and diversity of instream and riparian habitat, carbon inputs, dissolved oxygen levels, sediment transport and river morphology, and hydrological extremes, such as winter low flows and floods (Prowse, 2001a,b; Prowse and Culp, 2008). Within the Arctic, lake and river ice also permit the seasonal development of a suite of private and public transportation routes linking northern communities, thereby providing an inexpensive way to resupply remote resource industries and ready access for supporting traditional subsistence-based lifestyles, which depend on these lentic and lotic ecosystems (Nuttall et al., 2005; Furgal and Prowse, 2008; Prowse et al., 2009a,b).

The largest economic costs of lake and river ice are associated with the dramatic ice and flooding that accompany dynamic freeze-up and break-up events. Many Arctic communities were established at the confluence of rivers or where rivers enter lakes, and these sites are known to be highly susceptible to ice-jam formation. For example, the relocation of the administrative center from Aklavik, Canada, in the center of the Mackenzie River Delta, to Inuvik, higher up on the eastern banks of the Delta, was conducted to avoid ice-induced flooding. Although there are no summary figures for the costs of ice-jam flooding in the Arctic, the annual cost for North America is approximately 280 million USD (converted to 2009 values from Carlson, 1989; Gerard and Davar, 1995) and, for a single event in eastern Russia, the cost exceeded 124 million USD (2001, converted to 2009 values from Brakenridge et al., 2001). Damage from ice action and flooding also poses major economic costs for in-channel uses of northern rivers including bridges, pipelines and drilling platforms, transportation, and hydroelectric power generation. For example, designing around river-ice effects has been estimated to translate into losses of tens of millions of dollars per year for one northern hydroelectric power producer in Canada, further amounts also being spent just to monitor and mitigate ice problems (Burrell, 2008). In addition, there is the hazard to life associated with the risks of on-ice travel, ranging from commercial traffic on ice roads to the use of back-country lake- and river-ice networks for travel and access to traditional food sources by northern residents (e.g., Ford et al., 2008). More details on hydrological, ecological and socio-economic impacts are discussed in Section 6.5.

6.2. Sources of information

- Surface-based observations have provided much of the information regarding the response of lake and river ice to climate, the earliest long-term ice observations dating back to the 17th and 18th centuries.

- Over the past two decades, there appears to have been a significant decline in a number of ice observations, and there are fewer observations carried out at higher latitudes compared to more southerly locations.

- To extend the historical record, a number of other methods have been used, including isotopes, tree scars, and taxa buried in lake and pond sediments.

- For several years, remote sensing has been seen as the technology that would eventually supersede surface-based observations of river and lake ice, but this has not yet been realized in any circumpolar country.

- A majority of the historical trends about lake and river ice have been derived from surface-based observations. These are now in a state of decline, however, and remote-sensing approaches are emerging as an observational product that should increase the geographical range of coverage.

6.2.1. Surface-based observations

Surface-based observations have provided much of the information regarding the response of lake and river ice to climate. The earliest long-term ice observations on Arctic lakes and rivers date back to the 17th and 18th centuries. In Finland, for example, observations of ice cover break-up on the Tornionjoki River were initiated in 1693 (Kuusisto and Elo, 2000); in Sweden, on the Torneträsk River in 1700 and in Lake Mälaren in 1711; and in Russia, on the Neva River at the beginning of the 18th century. However, such long-term ice break-up records are available for a very limited number of sites. In general, most river- and lake-ice observational networks were established between the latter part of the 19th century and the beginning of the 20th century. During this period, regular observations were initiated on lakes and rivers of the United States, Canada, Russia, Scandinavia, and Western Europe, mostly under the responsibility of national hydrology and environmental agencies. The first observation programs involved the determination of basic lake- and river-ice regime characteristics (mainly dates of freeze-up or ice-on and break-up or ice-off). As the number of sites on the northern lakes and rivers of North America and Eurasia grew, the range of observed characteristics also increased. For example, dates of formation of first ice, complete freeze-over of the water body, initiation of ice melt, and when the water body becomes clear of ice also began to be recorded (Lenormand et al., 2002). Since the middle of the 20th century, national programs of ice thickness measurements have been carried out in most northern countries, some starting even earlier. In the former Soviet Union and in Canada, such measurements were initiated on rivers and lakes in the 1940s and 1950s, and were accompanied by measurements of the thickness of snow cover on ice (Vuglinsky and Gronskaya, 2006). In Finland, since the late 1970s, snow ice (white ice) thickness and snow thickness on ice have been measured in addition to the total thickness of the ice cover (Korhonen, 2006).

It is important to note that many countries use different methods of ice-regime observations. This relates especially to dates of ice formation, and to duration of freeze-up and break-up. For example, some countries document the initiation of break-up, while others register the date when the water body becomes completely ice-free. This is an important distinction when examining variability and trends in ice records from different sources, given that the entire break-up process can last for up to four weeks at a single site (Prowse et al., 2007a). Therefore, one of the primary objectives for improving instrumental records of lake- and river-ice regimes should be to standardize methods for measuring basic ice characteristics used in different countries.

Ice observations are usually kept in national archives and sometimes, after having been generalized, are published in various reports. In the past century, the data were most commonly published in annual

reference books. In addition to publications, national electronic hydrological and cryospheric databases containing ice observation data have been extensively developed over the past 20 to 25 years. The Global Lake and River Ice Phenology Database containing observations from many countries (until about 1996) is available through the National Snow and Ice Data Center (NSIDC) at the University of Colorado at Boulder, United States (Table 6.1) (Benson and Magnuson, 2007). This database contains freeze-up and break-up dates, as well as other ice-cover descriptive data for 748 lakes and rivers across the Northern Hemisphere. Of the 429 water bodies that have records longer than 19 years, 287 are in North America and 141 are in Eurasia; 170 have records longer than 50 years, and 28 have records longer than 100 years. A few have data prior to 1845. However, Table 6.1 shows that the number of stations decreases with latitude, particularly north of the Arctic Circle. Table 6.1 also includes those stations used by Magnuson et al. (2000) in an analysis of long-term records (see Section 6.4.2.1). In general, such data allow the analysis of broad spatial patterns and time series of freshwater-ice cover. The NSIDC has developed a web-based user interface that allows users to search the database and retrieve data according to various parameters. The interface includes a link to more general information about the lakes and rivers in the database (Benson and Magnuson, 2007).

Table 6.1. Number of lake and river observation sites north of specified latitudes contained in the current Global Lake and River Ice Phenology Database (GLRIPD) and those used by Magnuson et al. (2000).

	Number of stations north of 50° N	Number of stations north of 55° N
GLRIPD sites		
Total number of sites	465	317
Sites with post-1996 data	0	0
Sites with post-1990 data	152	110
Sites with post-1980 data	353	257
Sites with post-1970 data	381	273
Long-term sites (over 150 years) used by Magnuson et al. (2000)	9	6

In the near future (2012/13), surface-based river- and lake-ice thickness data will be available from the International Data Centre on the Hydrology of Lakes and Reservoirs (HYDROLARE), which is operated at the State Hydrological Institute (Russian Federation) under the auspices of the World Meteorological Organization (WMO). The objective of the center is the establishment, development, and regular updating of a global database on the hydrological regime of lakes and reservoirs, including river and lake ice data. The first tranche of data is expected to be received from WMO countries during 2010 for input to the database.

Surface-based observations were once the most important source of information regarding lake- and river-ice conditions. The declining state of these networks since the mid-1980s has led to serious geographical and temporal gaps for several lake and river ice parameters. For example, Figure 6.4 shows the precipitous decline evident during the latter part of the 20th century in observations recorded in Canadian, Russian, Swedish, and global databases. Although some of this decline is due to a reduction in the number of observing stations, some may also be due to a lack of reporting to national and international data centers. The two main reasons for the dramatic decline in the surface-based networks are the automation of meteorological stations near lake- and river-ice observation sites, and financial cutbacks (e.g., Lenormand et al., 2002). Another shortcoming of the databases is that they do not include ice-observation data (e.g., water level to bottom of ice-depth measurements, ice effects on water-level recordings and/or direct visual observations) that are routinely collected as part of national river hydrometric programs, and which can be excellent sources of information related to river-ice dynamics (see Section 6.5.1). Significant effort would be required to extract comparative information from the original archives (e.g., de Rham et al., 2008a; Goulding et al., 2009b; von de Wall et al., 2009).

Some attempts have been made to reverse this trend of declining observation programs. In Canada, for example, a number of ice-thickness monitoring sites have been re-established by the Canadian Ice Service

(CIS) as a contribution to the International Polar Year (IPY). In addition, at least two volunteer observational networks have recently been established in North America: IceWatch in Canada (www.naturewatch.ca/english/icewatch) and the Lake Ice and Snow Observatory Network (ALISON) in Alaska (www.gi.alaska.edu/alison). Various surface air-temperature indices are often employed to link trends in freshwater ice to the corresponding climate (see Section 6.4.2.2). These indices (e.g., degree-day calculations) can also be used to estimate phenological dates of break-up and freeze-up. This is particularly useful in regions where direct ice observations are not available and surface-temperature measurements from existing meteorological stations are incorporated to assess historical trends and/or projected future changes to ice phenology over specified regions.

Isotopes have also been used to index changes in lake and river ice. For example, vertical sectioning of river-ice cores and analysis of their oxygen-hydrogen isotopes has been used to provide an indirect record of the source of varying flow contributions (e.g., from lake water, groundwater, or direct precipitation) occurring during the complete period of winter ice growth (e.g., Gibson and Prowse, 2002). Analyses of sediment deposition records in riparian environments (e.g., ox-bow lakes), using a combination of isotopic and related geochemical analyses, have been used to interpret historical (multi-century) records of flood and related ice-jam events (e.g., Wolfe et al., 2006). Ice-related flood events on lakes and rivers have also been determined using ice-induced scars on trees (e.g., Bégin, 2000, 2001; Prowse and Culp, 2008). However, most of the isotopic and ice-scar data analyses have typically been conducted at single sites and at lower latitudes. Nevertheless, significant high-latitude work has been undertaken using changes in the taxa (i.e., siliceous algal and chitinous invertebrate remains) buried in lake and pond sediments to obtain long-term records of warming trends and historical changes in freshwater-ice cover. The strength of such inferences is based on the understanding that changes in ice-cover duration and, hence, longer growing seasons increase primary production and cause taxonomic shifts in algal and invertebrate communities. Since most of the analyses have been conducted on ponds and relatively shallow lakes, caution must be used in extrapolating results to larger lake systems. A synthesis of relevant paleo-information is provided in Section 6.4.1 to give a paleo-historical background to changes in Arctic freshwater-ice conditions.

6.2.2. Remote sensing

For several years, remote sensing has been seen as the technology that would eventually supersede surface-based observations of river and lake ice, at least for simple ice phenology and possibly for ice dynamics (e.g., break-up floods). Unfortunately, this transition has not yet been realized in any circumpolar country. However, considerable research has focused on evaluating the ability of different satellite remote-sensing datasets and related methods to derive freshwater-ice parameters on Arctic lakes and rivers. For example, the historical satellite archive from the Advanced Very High Resolution Radiometer (AVHRR) optical sensor has recently been used to assess trends in lake-ice phenology over 36 large lakes in Canada (six located in the Arctic; see Section 6.4.2.1) from 1984 to 2004 (Latifovic and Pouliot, 2007). Some studies have also shown the potential of AVHRR and the Moderate Resolution Imaging Spectroradiometer (MODIS) to monitor break-up dates on very large northern-hemisphere rivers (e.g., Pavelsky and Smith, 2004). The spatial resolution of most satellite sensors providing high (daily) temporal resolution has simply been too coarse for monitoring the majority of river-ice parameters. Another problem with optical data is that they suffer from atmospheric interference (e.g., clouds) and polar darkness, which limits ice-phenology parameter retrievals during crucial times (especially during freeze-up).

The microwave region of the electromagnetic spectrum is perhaps more useful for the monitoring and study of lake-ice phenological processes because of its all-weather and polar-darkness imaging capability. For example, passive microwave data from the Defense Meteorological Satellite Program Special Sensor Microwave/Imager (SSM/I) 85 GHz brightness temperature channel has been used to detect spatial phenology changes over Great Slave Lake, Northwest Territories, Canada (Schertzer et al., 2003). However, the 85 GHz channel is susceptible to considerable atmospheric interference, and its spatial resolution is still relatively coarse (12.5 km), which can lead to large brightness temperature differences

between water and land (Cavalieri et al., 1999). Nonetheless, it has been shown that ice regimes in the main section of lakes (i.e., away from shore that can contaminate the coarse-resolution pixels) can be monitored when passive microwave is used synergistically with radar altimeter data.

Kouraev et al. (2007a) developed an approach to monitor ice cover on Lake Baikal by combining satellite altimetry (TOPEX/Poseidon, Jason-1, ENVISAT, and Geosat Follow-On) with SSM/I data. The new approach allowed a reconstruction of the ice regime in the southern as well as the central and northern parts of the lake from 1992 to 2004 (Kouraev et al., 2007b). Kouraev et al. (2008) further demonstrated that the combination of passive and active microwave (altimeter) data could be extended to other large lakes and interior seas of the Eurasian continent – the Caspian and Aral Seas, Lake Ladoga and Lake Onega – and possibly to the large northern lakes of Canada, such as Great Slave Lake and Great Bear Lake.

More recently, Howell et al. (2009) developed algorithms to detect melt onset, water clear of ice, and freeze onset dates on Great Bear Lake and Great Slave Lake, Canada, using moderate resolution (~4 km) spaceborne scatterometer data from QuikSCAT (Quick Scatterometer satellite). Some researchers have also used high-resolution (30 m) ERS-1 or ERS-2 (European Remote Sensing satellite) synthetic aperture radar (SAR) imagery to monitor ice formation, the thickening of ice cover, and freezing to the bottom of shallow (e.g., less than ~2 m) Arctic and sub-Arctic lakes in Alaska (Jeffries et al., 1994; Morris et al., 1995) and northern Manitoba, Canada (Duguay et al., 1999; Duguay and Lafleur, 2003). Duguay et al. (2002) utilized RADARSAT-1 SAR imagery for monitoring ice growth and decay and related processes of shallow sub-Arctic (tundra and forest) lakes in northern Manitoba, Canada. Nolan et al. (2003) used ERS-2 and RADARSAT-1 data, together with a numerical ice model, to gain an understanding of the ice dynamics on Lake El'gygytgyn, northeastern Siberia. The value of higher resolution RADARSAT imagery (8 m) has also been demonstrated for the mapping of river-ice types (Weber et al., 2003).

Ice thickness has been estimated with some success through the synergistic use of optical and SAR data on shallow Arctic lakes (e.g., Duguay and Lafleur, 2003). Coarse resolution Advanced Microwave Scanning Radiometer - Earth Observing System (AMSR-E) data acquired in the 6 to 18 GHz frequency range have also been shown to be suitable for estimating ice thickness on very large lakes (Kang et al., 2010). In addition, recent work has demonstrated the possibility of determining total ice thickness of snow and ice on medium to large lakes from snow-surface elevation data acquired by spaceborne radar and laser altimeters (e.g., Kouraev et al., 2008). However, due to the failure of one of three lasers of the Geoscience Laser Altimeter System (GLAS) instrument shortly after the launch of the Ice Cloud and Land Elevation Satellite (ICESat), measurements are very limited temporally. The replacement ICESat-2, scheduled for launch in 2015, will provide an opportunity to further explore the potential of laser altimetry for estimating snow and ice thickness on northern lakes.

Operational ice monitoring from large lakes is conducted by the Canadian Ice Service (CIS) and the National Oceanic and Atmospheric Administration (NOAA) National Environmental Satellite, Data, and Information Service (NESDIS) in the United States. The CIS began operational weekly monitoring of ice extent on large lakes in 1995 using NOAA AVHRR (1.1 km) and RADARSAT ScanSAR (100 m) imagery to meet the Canadian Meteorological Centre's needs for lake-ice coverage in numerical weather prediction models. The amount of ice on each lake (in tenths) is determined by visual inspection of AVHRR and RADARSAT imagery. The program started with 34 lakes (in 1995) and has now reached 136 lakes (mostly in Canada with a few in the United States). Using this dataset, it is possible to derive dates of complete freeze-over and when water is clear of ice with an accuracy of ±1 week, a temporal resolution not sufficient for climate change studies. The accuracy of the information in the CIS lake-ice coverage database depends on the amount of cloud cover over a particular lake (NOAA AVHRR imagery is cloud dependent) and the frequency of RADARSAT coverage. In 1997, NESDIS started to generate a daily snow and ice product at a resolution of about 24 km with the Interactive Multisensor Snow and Ice Mapping System (IMS). The IMS incorporates a wide variety of satellite imagery (AVHRR, GOES, SSM/I, etc.) as well as derived mapped products (e.g., U.S. Air Force Snow and Ice Analysis) and surface

observations. The coarse resolution of the 24-km product allowed mapping of ice extent on only the largest lakes of the Northern Hemisphere. Since February 2004, a 4-km resolution product has become available, which permits determination of ice extent on lakes of this resolution or greater on a daily basis.

Significant advances have taken place on the research front and in the development of operational products, and capabilities for ice monitoring of many lakes and rivers of the Arctic at high spatial (50 to 100 m) and temporal resolution (daily) are foreseen in the not too distant future. The Sentinel satellite missions of the European Space Agency (ESA, 2011 to 2012) and the RADARSAT constellation (2014 to 2016) of the Canadian Space Agency (CSA) will provide the technology needed to achieve the goals set by the Global Climate Observing System (GCOS) regarding the accuracy for determining freeze-up and break-up dates (± 1 to 2 days).

6.3. Role in the climate system

6.3.1. Effects on local climate and large-scale climatic feedbacks

- Lake and river ice need to be considered in climate modeling given the large fraction of the northern high-latitude, sub-Arctic and tundra landscape that they occupy.

- The degree to which the atmosphere is influenced depends on several factors including the magnitude, timing, location and duration of ice cover.

- A significant shortening of the freshwater ice duration period can affect local, regional, and even larger-scale climate over the Arctic.

- Ice-cover duration and, to a lesser degree, thickness have been shown to play roles in the annual energy and water balance of large Arctic river basins, such as the Mackenzie, Canada.

In the Arctic, the meteorological and climatological effects of freshwater ice within the terrestrial landscape are mostly confined to the local scale, with the greatest effects produced by ice cover on large lakes. The presence or absence of ice directly modifies atmospheric heating of lakes and rivers through changes in latent and sensible heat fluxes, absorption and reflection of short-wave radiation, and long-wave radiation emissivity of the surface. This in turn affects local climatic factors, primarily surface-air temperature, but also precipitation, evaporation, and low-level cloud cover. The degree to which the atmosphere is influenced depends on several factors including the magnitude, timing, location, and duration of ice cover (e.g., Rouse et al., 2005). Ice-covered freshwater bodies of various sizes represent a considerable fraction of the high-latitude, sub-Arctic and tundra landscape (see Section 6.1.3), and as such need to be considered in any type of climate modeling of Arctic terrestrial environments (e.g., Samuelsson et al., 2008). For example, the surface area of lakes in the Mackenzie River Basin of northwestern Canada is about 144 000 km^2 or 8% of the total area (Bussières, 2002), while in the northern Hudson Bay Lowlands, Bello and Smith (1990) estimated the lake coverage to be 41% of the landscape. Other high-latitude regions, such as Alaska, northern Scandinavia, and northern Russia, also exhibit a substantial areal coverage by lakes (e.g., Ljungemyr et al., 1996). The lateral dimensions of large rivers can also approach those of small lakes, and the scale of their climatic impacts is likely to be comparable. The greatest effects are expected for the deltas of large river systems in which the network of tributary channels and interconnected ponds and lakes encompasses a vast region of the landscape (see Section 6.5.2.3). As a result, a significant shortening of the freshwater-ice duration period can have profound impacts on local, regional, and even large-scale climate over the Arctic (e.g., Rouse et al., 2005).

Climatic impacts from terrestrial water bodies are fairly well documented during open-water periods, but less is known regarding ice-covered conditions, with most information focusing on transitions in spring and autumn. During spring, an ice-covered water body has a significantly different energy balance to that of snow-free land, with nearshore zones usually characterized by cooler spring air temperatures. Studies

have shown that the inclusion of lakes in numerical weather-prediction models can produce a springtime cooling of near-surface air temperature of up to several degrees Celsius in the vicinity of lakes when they are ice-covered (e.g., Mackay et al., 2009). For large river deltas, ice break-up has been determined as the critical event impacting the thermal regime of local to regional climate. Arrival of the break-up front causes a rapid clearing of snow and ice leading to a dramatic reduction in surface albedo. During a dynamic break-up, local air temperatures have been found to rise rapidly, by as much as 5 °C, whereas thermal break-up (with less flooding) had minimal effects on surface temperature. Biological impacts can also be highly visible, since slight increases in radiative warming could raise air temperatures above a critical threshold value that stimulates budding and early-spring plant growth (Prowse, 2000). These processes are scale dependent and, therefore, ecological impacts resulting from changes in ice cover on small rivers and streams would be primarily restricted to very local effects (e.g., to the riparian zone; Prowse and Gridley, 1993).

Regional to large-scale climatic feedbacks from freshwater bodies are mainly associated with big lakes and, to some degree, large river-delta systems. With regard to the former, there have only been a few modeling and observational studies that have examined the impact of lakes on regional to large-scale climate. As expected, results revealed a significant enhancement of latent heat flux and suppression of sensible heat flux when freshwater lakes are present (e.g., Nagarajan et al., 2004; Rouse et al., 2005). One of the most comprehensive analyses related to freshwater-climate links in the Arctic was undertaken in the Mackenzie GEWEX (Global Energy and Water Cycle Experiment) Study that comprised a series of large-scale hydrological and related atmospheric and land-atmosphere studies of the Mackenzie River Basin. Results demonstrated that ice-cover duration and, to a lesser degree, ice thickness play significant roles in the annual energy and water balance of large river basins such as the Mackenzie River Basin. This is mainly because ice cover prohibits evaporative exchange with the atmosphere for several months of the year. For large lakes, the evaporative and sensible heat fluxes reach their maxima during autumn and early winter mainly due to the greater frequency of cold air masses over relatively warm lake surfaces (Rouse et al., 2003). Such heat fluxes remain large as long as there is open water. For example, significant evaporation from Great Slave Lake occurs over a period ranging from six to seven months (Blanken et al., 2007). Moisture from this evaporation can be an important source for subsequent precipitation within the Mackenzie River Basin (e.g., Szeto, 2002). Unfortunately, precipitation measurements are typically so sparse that the contribution of lakes cannot be quantified from direct measurements. Nonetheless, large-lake evaporation in autumn and early winter is known to trigger frequent downwind snow squalls, as seen from passive microwave satellite images of the Great Slave Lake and Great Bear Lake region (i.e., similar to the Canadian Laurentian Great Lakes). However, no substantive studies have yet been pursued on this feature (Walker et al., 2000). It should be noted that many of the findings from the Mackenzie GEWEX Study are also applicable to other high-latitude regions of the world (Rouse et al., 2007).

Samuelsson et al. (2010) used a numerical regional climate model to determine how the presence of lakes affected regional to large-scale climate over Europe. Results showed that the incorporation of the observed lake network into the model resulted in a warming effect on European climate for all periods except mid-winter to mid-spring when ice cover was present. The warming was particularly pronounced during autumn and winter, and over regions that have a high concentration of lakes. Locally, the presence of lakes enhanced convective precipitation by 20% to 40% in late summer and early autumn; however, this precipitation was reduced by more than 70% during early summer.

High-latitude lakes are more prone to be affected by climate variability and change than the surrounding landscapes, mainly due to their strong absorption of solar radiation during the ice-free period. Therefore, shorter ice duration will act to increase annual net radiation, heat storage, and evaporative and sensible heat fluxes (Rouse et al., 2007). Such increases in net radiation will add to the heat content of the lake, as was shown for Great Slave Lake in the Mackenzie River Basin during the unusually warm year of 1998 when the ice-free period was about 40 days longer than normal (Schertzer et al., 2007). The larger heat storage will subsequently increase open-water evaporation and the overall sensible heat flux, particularly

during autumn and early winter (for large lakes), which can then augment lake-effect snowfall by depositing more snow on downwind terrestrial locations.

6.3.2. Climatic controls

- In addition to the various climatic variables that control surface heat fluxes, freshwater-ice processes and phenologies are also controlled by landscape hydrology.
- Through modifications of terrestrial hydrology, climate also plays an indirect role in affecting freshwater-ice phenologies.
- In the case of lakes, runoff inputs directly influence freeze-up and break-up through the direct addition of heat, the modification of surface albedo, and/or the creation of intra-lake currents and mixing.
- The entire ice regime of rivers, including phenology, ice-cover thickness, and ice-jamming processes, is strongly influenced by hydrological variables.

6.3.2.1. Direct controls

The formation, growth, decay, and break-up processes of the ice cover differ markedly between lakes and rivers, but all are influenced by similar climatic variables that control surface heat fluxes, such as solar radiation, air temperature, humidity, precipitation, wind speed, and cloud cover, etc. (Ashton, 1986; Walsh et al., 2005). River and lake size and depth have a strong effect on the timing of freeze-up, owing to the heat content of large water bodies. Freeze-up dates can vary by as much as two months between small and large lakes in the same region; however, size has no similar effect on break-up, which may only vary by about two weeks (Rouse et al., 2007). Air-temperature indices are often used for approximate prediction of ice-cover phenology and thickness (Borshch et al., 2001; Vuglinsky and Gronskaya, 2006), but the resulting correlations are site-specific as they tacitly assume that all other relevant climatic variables are either constant or singly related to air temperature (Beltaos and Prowse, 2009). Physically based process models quantify the synergistic effects of the relevant climatic variables, but may require specialized input data that are not always available (Duguay et al., 2003; Liu et al., 2006; Saloranta and Andersen, 2007). Prediction of break-up timing is hampered by knowledge gaps with respect to climatic effects on the composition of the ice cover, which partly controls internal decay by absorption of solar radiation (e.g., Hicks et al., 2008). Not only does ice composition differ between rivers and lakes, it can also change during the ice season, thus altering absorption characteristics (Walsh et al., 2005). A key mechanism in promoting change in cover composition is the winter loading of snowfall; if sufficient snow accumulates to depress the ice cover below the hydrostatic water level, surface flooding (or 'slushing') will result and with subsequent refreezing will become highly reflective surface layers of white ice (see Section 6.1.3).

6.3.2.2. Indirect controls

6.3.2.2.1. Lake ice

In addition to direct atmospheric budget controls of lake ice, mechanical action of wind and landscape hydrology can also significantly affect ice timing and duration. Inflow from streams or land runoff can influence break-up, affecting the phenology by adding heat inputs, creating currents within the lake, or both. In the case of small lakes, for example, the spreading of spring meltwater on the surface can play an important role in decreasing surface albedo and, hence, can indirectly advance the radiation decay of the ice cover (Woo and Heron, 1989; Grenfell and Perovich, 2004). For large lakes, such processes are primarily limited to the margins where meltwater moats typically develop at the early stages of break-up or by the inflow of warmer water from rivers, which accelerates the break-up process. For example, MODIS satellite imagery over Great Slave Lake from 2004 to 2006 (Figure 6.5) shows a crack forming in the ice cover during break-up near the large inflow from the Slave River. The largest indirect hydrological

control relates to the volume and temperature of surface water and groundwater as these control the open-water heat budgets of lakes. As noted in Section 6.3.2, the heat storage of lakes greatly affects the timing of autumn freeze-up (e.g., Blanken et al., 2007). The duration of the ice-free period of summer warming is particularly important for stratified lakes where an earlier start to the stratified season can significantly increase the period over which a lake warms, leading to a greater increase in mean summer lake temperatures than would be expected simply from summer air temperatures (Austin and Colman, 2007).

6.3.2.2.2. River ice

In rivers, the entire ice regime, including phenology, ice-cover thickness, and ice-jamming processes, is strongly influenced by hydrological variables, such as the flow regime and the base level, both of which are partly controlled by climate (Beltaos, 2008b; Beltaos and Burrell, 2008). For example, high autumn discharge tends to delay freeze-up, while high spring discharge tends to advance break-up and enhance the severity of ensuing ice jams and related flooding. The effects of discharge on river-ice processes arise from its controlling influence on flow hydrodynamics, including such variables as depth, velocity, erosional capacity, forces applied on the ice cover, and water surface slope. In terms of water surface slope, an additional influencing factor is the elevation of the water body into which a river may drain, such as another river, a lake, or the ocean. The effect of changes in the ocean is discussed in Section 6.5.1.1. Complexity is further amplified by antecedent conditions (freeze-up levels and ice-cover thickness) playing a significant role in the timing and evolution of break-up events, which often generate extreme ice jams and floods. Although existing process-based modeling capability (Liu et al., 2006) can largely account for many direct and indirect climatic effects, breakup-related phenomena have not yet been adequately quantified (Beltaos, 2008c; see also Section 6.5.1.1).

6.4. Changes: past and future

6.4.1. Paleo-historical changes

- Arctic lakes act as natural recorders of past variation in ice cover because their biological activity is closely coupled to the physical properties of stratification and ice cover, and because their fossil record has been little disturbed by human activities.

- All proxies of lake primary production remained low throughout the middle and late Holocene, in agreement with cold prevailing conditions and prolonged ice cover. Pronounced increases in whole-lake productivity, however, occurred during the warm period of the Holocene Thermal Maximum, consistent with decreased ice cover and its influences on in-lake dynamics.

- Ice-cover duration became equally short or shorter than it is now during the Medieval Warm Period. Arctic lakes were also more productive during the Medieval Warm Period than in the subsequent Little Ice Age – the dominant control on productivity being the duration of the ice-free season.

- During the Little Ice Age, ice-cover duration lengthened and probably led to anoxic conditions in lake bottom waters in some parts of the Arctic.

- Since the beginning of the 19th century, some Arctic lakes have experienced longer ice-free seasons, or accelerated ice melt in the case of some that have been perennially ice-covered.

- Inferences from some Arctic lake records suggest that many lakes may have crossed an important ecological threshold as a result of recent warming.

To assess the significance of changes in modern properties of lake and river ice, it is essential to place recent observations in a longer-term context. However, long-term monitoring data about the freezing and

thawing of Arctic lakes are scarce. Inferences into past freshwater changes can be made by examining samples that record environmental conditions in the paleolimnological record. Paleolimnological analyses can provide annual to millennial time series of the composition of lentic biological communities, biogeochemical processes, and changes in lake physical conditions, which can be used for analyses of climate effects on lakes (Battarbee et al., 2005). Detailed analysis of lake sediments can identify the mechanisms by which climate affects lakes, as well as how climatic variability may interact with other regulatory processes over timescales inaccessible by standard limnological approaches (Leavitt et al., 2009).

Remote Arctic lakes have high potential as natural monitors and recorders of past variation in ice cover because their biological activity is closely coupled to the physical properties of stratification and ice cover, and because their fossil record has been little disturbed by human activities. However, since most paleo-analyses have been conducted on ponds and relatively shallow lakes, the data must be used circumspectly to infer changes on larger lake systems. The following sections summarize evidence from proxy climate and paleolimnological data with regard to the nature and causes of past variability in Arctic climate and lake ice conditions over the Holocene interglacial period and the more recent Anthropocene period.

6.4.1.1. Holocene variability in lake ice conditions

Many physical, chemical, and biological changes occur in Arctic lakes that are either directly or indirectly affected by snow and ice cover (see Section 6.5). Recently, attempts have been made to directly link changes in lake biological communities (mainly diatoms, which are important primary producers in Arctic lakes and preserve well in sediments) to the measured or empirically modeled duration of ice cover using extensive modern calibration datasets together with careful statistical analyses (Korhola, 2007). The central ecological argument behind the approach is that an extensive layer of ice and snow would dramatically affect limnological conditions, resulting in marked changes in biological communities (Douglas and Smol, 1999). In years with cold summers, with more extensive ice cover, diatoms from aerophilous (oxygen supplied) and shallow habitats will dominate. Conversely, during warmer years, there is less extensive ice cover and taxa characteristic of deeper-water substrates and planktonic habitats will increase in abundance relative to the shallow-water benthic taxa (Smol and Douglas, 2007a). Eventually, in deeper lakes, thermal stratification may also occur (or be prolonged) during the summer months, further altering the composition of the diatom communities. Increased stratification has been shown to favor small-celled diatoms with a high surface area to volume ratio (Rühland et al., 2008; Winder et al., 2009). Using a pan-European dataset of species abundance for 252 diatom taxa in 459 mountain and sub-Arctic lakes, Thompson et al. (2005) developed a transfer function to infer duration of ice cover based on different diatom assemblage species compositions. These are now being developed for application to lakes at higher latitudes.

In the absence of suitable transfer functions, less precise ice-cover reconstructions are commonly based on changes in species assemblage and the known ecological and life-history characteristics of the organisms. However, if ice and snow cover change in a lake, there should be changes not only in the relative frequencies of different taxa, but also in overall primary production in the lake. Moreover, the length of ice cover can also drive changes in key limnological variables, including nutrient levels, mixing regimes, gas exchange and, in poorly buffered sites, fluctuations in lake-water pH that can be inferred from biotic and abiotic proxies (Douglas and Smol, 1999). The importance of each of these factors is site-specific and varies over time; thus, multiple hypotheses explaining past changes in proxies are often warranted.

Numerous paleolimnological studies exist for Arctic lakes covering the entire Holocene, yet studies focusing explicitly on ice-cover duration are scarce and limited to small and shallow systems. Michelutti et al. (2007) used many paleolimnological techniques including reflectance spectroscopy and diatoms to infer past trends in primary production and lake-water pH from the sediments of a small lake on Baffin Island in the Canadian Arctic. Pronounced increases in whole-lake productivity and lake-water pH were noted during the warm periods of the Holocene Thermal Maximum (~10 000 to 8000 BP) and during the

recent climatic warming, consistent with the hypothesis of decreased ice cover and its influences on in-lake dynamics, particularly dissolved inorganic carbon. All proxies of lake primary production remained low throughout the middle and late Holocene, in agreement with cold prevailing conditions and prolonged ice cover. However, maximum recent values of their reconstructed parameters were again either directly comparable to or exceeded values attained during the early Holocene. There are many examples of lakes on Baffin Island that show similar developmental characteristics (see Wolfe and Smith, 2004).

Almost parallel development regarding ice-cover duration was observed by Cremer et al. (2004) in a weakly thermally stratified lake in the Ural Mountains of northern Russia. Using diatoms and a range of geochemical indices (e.g., total carbon, total organic carbon, total sulfur), they concluded that ice-free conditions prevailed in the lake during the Holocene Thermal Maximum and were characterized by high biological productivity and strong growth of planktonic diatoms. Since then, the duration of the open-water period gradually decreased in concert with Neoglacial cooling, followed by a return to conditions similar to those during the Holocene Thermal Maximum beginning a few hundred years ago.

More detailed studies have increasingly been conducted, some based on annually laminated sediment sequences, covering the past few millennia with a focus on finer-scale climate oscillations. Tomkins et al. (2009) combined the results from a varve (a layer of sediment deposited in a lake in a single year) chronology that included records of a novel sediment feature ('sedimentary pellets'), which they interpreted as an ice-cover indicator, to construct a 1000-year proxy record of ice-cover extent and dynamics on a perennially ice-covered, High Arctic lake on Ellesmere Island. Sedimentary pellet frequency from multiple sediment cores suggested that the most notable period of reduced ice cover was from about 1891 to present. Another period of ice-cover mobility was suggested from about 1582 to 1774, while persistent ice cover is inferred during the 1800s and prior to 1582. The proxy ice-cover record was found to correspond well with regional proxy temperature and paleoecological records, especially during the 1800s and 1900s.

Over approximately the past millennium, evidence indicates that lakes in the Arctic were more productive during the Medieval Warm Period and less productive during the subsequent cooler Little Ice Age – the dominant control on productivity being the duration of the open-water season. Tiljander et al. (2003) discovered a suite of organic-rich sediments from a varved lake in Finland with dates including the Medieval Warm Period (980 to 1250). During this time interval, less mineral material accumulated on the lake bottom than at any other time in the 3000-year sequence analyzed, suggesting a prolonged open-water period and negligible winter snow cover. In contrast, extremely thin varves and low sediment accumulation prevailed during the Little Ice Age in many lakes throughout the Arctic, suggesting an extensive ice cover and a short growing season (e.g., Lamoureux and Bradley, 1996; Hughen et al., 2000; Moore et al., 2001; Smith et al., 2004; Besonen et al., 2008; Bird et al., 2009). Prolonged ice cover also led to widespread anoxic conditions in lake bottom waters. For example, it has been reported that Lower Murray Lake in northern Ellesmere Island suffered widespread anoxic conditions from about 1700 to the mid-19th century, when the coldest episode of the late Holocene occurred in that region (Besonen et al., 2008).

6.4.1.2. Anthropocene variability in lake ice conditions

There are many examples of profound aquatic responses to recent Arctic warming. For example, Douglas et al. (1994) applied modern paleolimnological techniques to study the environmental history of a series of ponds on Ellesmere Island in the Canadian High Arctic. Their sediment profile results showed that, following several millennia of relative ecological stability, episodes of nearly complete species turnover in diatom taxa occurred, beginning in the 19th century. The species changes detected showed a proliferation of littoral taxa, indicating more complex periphytic diatom communities, as well as increases in the percentages of moss epiphytes, all of which require a longer ice-free season to develop in the harsh Arctic environment.

Other paleolimnological studies, using similar procedures, have also documented pronounced ecosystem shifts in the circumpolar Arctic, based not only on diatoms but also on other taxa representing different levels in the food web (e.g., Korhola et al., 2002; Sorvari et al., 2002; Quinlan et al., 2005; Rühland and Smol, 2005; Solovieva et al., 2005; Michelutti et al., 2006). The critical controlling role of ice cover in producing such shifts is evident in the results of a comparative sedimentary study of two lakes from northern Ellesmere Island with nearly identical limnological features, but different local climate and, therefore, degree of ice coverage (Keatley et al., 2008). The lakes recorded strikingly different changes in diatom assemblage, with the ice-covered lake tracking little ecological change, whereas the less-shaded lake recorded changes in its recent sediments consistent with warming.

Further evidence of accelerated lake-ice melt comes from lakes that had been perennially ice-covered or characterized by extended periods of ice cover. For example, in the rarely ice-free Sawtooth Lake situated in the Sawtooth Mountain Range of central Ellesmere Island, Perren et al. (2003) could not identify diatoms or chrysophyte cysts in sediment corresponding to the past 2500 years except since the 1920s when a rapid colonization of relatively diverse diatom algal flora occurred. The recent nature of this assemblage suggests a decrease in ice cover and a concomitant increase in light and nutrient availability for algal production over the past about 80+ years. Similarly, in Upper Dumbell Lake on northern Ellesmere Island, very few diatoms were recorded in sediments before about 1950, presumably due to harsh ice conditions (Doubleday et al., 1995). Only with accelerated warming over the past few decades did enough of the shallow-water moat thaw in summer to allow small, benthic taxa to increase in abundance. Still farther north, on Ward Hunt Island, Antoniades et al. (2007) recorded a similar appearance of diatoms, as well as dramatic increases in production (recorded by fossil algal pigments), in the recent sediments of Canada's northernmost lake (83°05´ N, 74°10´ W; Figure 6.6). The period of relative stability throughout much of the sediment core prior to the diatom and pigment shifts suggests that the lake's biology has been affected more by environmental change during the past two centuries than at any point during the preceding eight millennia (Antoniades et al., 2007).

In an attempt to synthesize the circumpolar paleolimnological records, Smol et al. (2005) compiled a dataset of 55 biostratigraphic profiles and estimated the amount of compositional change or turnover between the sediment samples at each site over the same time period (~150 years) using numerical procedures. They made the following conclusions: areas of the Arctic that were expected to have warmed the most also showed the greatest degree of compositional change; ecological change had occurred at several trophic levels; and the ecological characteristics of the species involved indicated that changes were driven primarily by climate warming. However, while most freshwater ecosystems show signs of pronounced change associated with warming in the High Arctic, similar changes are not yet detectable in lakes and ponds of northern Québec and Labrador (Pienitz et al., 2004). This remarkable regional stability at timescales of decades to hundreds of years is consistent with decadal observational (e.g., Serreze et al., 2000) and tree-ring (D'Arrigo et al., 2003) data that reveal climatic stability or even slight cooling over the western subpolar North Atlantic and adjoining land areas of eastern sub-Arctic Canada. It also suggests that northern Québec and Labrador may experience less short-term or delayed climate change relative to other Arctic regions and that they could be the ultimate bellwether of large-scale circumpolar change (Pienitz et al., 2004).

More recently, a meta-analysis of fossil diatom records from more than 200 lakes in the Northern Hemisphere revealed spatially structured but temporally coherent changes in species composition since the 19th century (Rühland et al., 2008). By comparing only a subset of highly resolved fossil time series with 30 to 100 years of environmental monitoring, and selecting only lakes in undisturbed catchments, Rühland and co-workers demonstrated that reorganization of community composition is correlated with change in ice cover and related limnological conditions (mixing and light regimes), rather than change in the influx of nutrients. They further inferred that many Arctic lakes may have crossed an important ecological threshold as a result of unprecedented atmospheric warming.

6.4.2. Trends of the instrumental period and linkage with climate

- Long-term records for the Northern Hemisphere (1846 to 1995) indicate the following (although only one site was located north of the Arctic Circle):
  - freeze-up has become later (average +6.3 d/100 y)
  - break-up has become earlier (average -5.8 d/100 y)
  - ice duration has decreased (average 12.1 d/100 y).

- A subsequent analysis of a much smaller subset of only lakes from the same archive with records to 2004/05 indicates the following further changes:
  - later freeze-up (+10.7 d/100 y)
  - earlier break-up (-8.8 d/100 y)
  - reduction in average ice duration (19.5 d/100 y)
  - strong and consistent changes after the mid-1990s probably influenced this rate change but different sample sizes and/or periods of record may also have played a role.

- Many shorter-term regional studies have also been conducted, but they exhibit appreciable spatial and interdecadal variability. Much of this is likely to be due to analytical differences in the number and location of station data and the various record lengths used for trend calculations.

- Most trends toward shorter freshwater ice duration over much of the circumpolar North closely correspond to increasing air temperature trends and, more specifically, timing of the 0 °C isotherm.

- Broad spatial patterns in ice trends have also been linked to major atmospheric circulation patterns, different phases of which can cause contrasting ice conditions (e.g., shorter versus longer ice duration) across individual continents and between opposite sides of the circumpolar North.

- Some important south-north contrasts, however, have also been identified in freshwater ice trends. Examples from Scandinavia show more pronounced change (later freeze-up and earlier break-up) occurring in southern than in northern lakes, perhaps indicating greater sensitivity to warming at the more temperate latitudes.

- Contrasting results, albeit involving a more recent period, have been noted for south-north regions of Canada based on records obtained by remote sensing. The degree to which this reflects the effects of either more recent or higher-latitude warming or a combination of both is unclear.

- Large-scale, comprehensive records of river- and lake-ice thickness are rare. One dataset compiled for Canada over the past 50 years does not reveal any obvious trends over the latter part of the 20th century. However, reductions in maximum ice thickness have been observed on nearly all rivers and lakes within Arctic Russia over the past two decades compared to the previous 30-year period.

6.4.2.1. Trends in instrumental records

Historical evaluations of changes in freshwater ice conditions over the circumpolar Arctic have proven difficult to synthesize owing to a variety of issues, including the limited number of detailed observations; different observational periods of record; varying *in situ* observational methodologies and/or phenological definitions; and changes in observation methods, such as from *in situ* to remote sensing (see also Section 6.2). As a result, most trend analyses have focused on relatively simple characteristics that are easy to glean from most agency records, such as the timing of autumn freeze-up and spring break-up, ice-cover duration, and ice thickness. The majority of these have been summarized by Walsh et al. (2005). They range from long-term records (~150 years or more) from a small set of observation sites around the Northern Hemisphere (Magnuson et al., 2000; Table 6.1) to several significant shorter-term (100 years or less) regional analyses over Arctic and sub-Arctic Russia (Ginzburg et al., 1992; Soldatova, 1993; Smith,

2000), northern Scandinavia (Zachrisson, 1989; Kuusisto and Elo, 2000), and northern areas of North America (Jasek, 1999; Sagarin and Micheli, 2001; Zhang et al., 2001). A detailed review of historical trends in the ice phenology of northern rivers was provided by Beltaos and Prowse (2009).

The long-term records originally analyzed by Magnuson et al. (2000) focused on the period 1846 to 1995, and included 39 time series of freeze-up or break-up from 26 sites (5 rivers, 19 lakes) with records of more than 100 years during this period. Only one site was north of the Arctic Circle (Table 6.1), reflecting the lack of high-latitude long-term observation sites. Overall, 38 of the 39 time series showed either later freeze-up (15 sites averaging +6.3 d/100 y) or earlier break-up (24 sites averaging -5.8 d/100 y), thus resulting in an average reduction in ice duration of 12.1 d/100 y[1].

In a subsequent analysis (by B.J. Benson and J.J. Magnuson, reported by Koç et al., 2009) of a smaller set of only northern-hemisphere lakes (9 sites for freeze-up and 17 for break-up) for the winters 1855/56 to 2004/05 from the same GLRIP database (Figure 6.7), the rate of change in both events is noted to increase: from +6.3 to +10.7 d/100 y for freeze-up and -5.8 to -8.8 d/100 y for break-up, thereby further reducing average ice duration by 12.1 to 19.5 d/100 y. It is unknown how much of this increased rate of change results from different sample sizes or periods of record. However, some is certainly due to the strong and consistent changes in timing that are evident after the mid-1990s (Figure 6.7), which is also the end of the previous period of analysis by Magnuson et al. (2000).

Among most of the shorter-term regional studies noted above, which typically contained far more sites than those used in the long-term evaluation by Magnuson et al. (2000), there is appreciable spatial and interdecadal variability. Much of this is due to analytical differences in the number and location of station data, and the various record lengths used for trend calculations. For example, river-ice trends in North America for the latter half of the 20th century exhibit a major spatial distinction between western and eastern regions, with the west showing significant and the east, small or insignificant trends toward earlier break-up (Walsh et al., 2005). Similar spatial trends were determined by Duguay et al. (2006) for Canadian lake-ice break-up. For Eurasia, Vuglinsky and Gronskaya (2006) found that average freshwater-ice duration decreased between two 20-year periods (1950–1979 and 1980–2000) by 2 to 10 days for lakes and rivers in European Russia, and by 3 to 7 days for rivers and 4 to 14 days for lakes in Asian parts of Russia. The reductions were attributed to earlier spring break-up and later autumn freeze-up.

Some important south-north contrasts have also been identified in freshwater-ice trends. For example, an investigation of Finnish lakes (with the longest time series from the early 19th century) showed that, with the exception of the far North, ice-cover duration has become significantly shorter due to later freeze-up and earlier break-up (Korhonen, 2006). Similarly, findings from Sweden for the period 1961 to 1990 show a rapid advancement in lake ice break-up of -0.92 d/y at southern latitudes, whereas the rate was significantly less at only -0.25 d/y for Arctic lakes (Weyhenmeyer et al., 2005). Contrasting results, albeit involving a more recent period, have been noted for south-north regions of Canada. Based on AVHRR imagery, Latifovic and Pouliot (2007) extended *in situ* records for 36 Canadian lakes for the 1950s to 2004 and developed new records for six high-latitude lakes from 1985 to 2004. Similar to most of the studies noted above, the majority of the sites showed earlier break-up (averaging -0.18 d/y) and delayed freeze-up (averaging +0.12 d/y) dates for the entire period. For the more recent period from 1970 to 2004, the rates increased to an average of -0.23 d/y and +0.16 d/y, respectively. As noted by Prowse and Brown (2010), however, the most rapid rates of change occurred in the six high-latitude lakes (primarily on the Canadian Archipelago) for the even more recent period of 1985 to 2004. For these lakes, changes in timing associated with earlier break-up and later freeze-up averaged -0.99 d/y and +0.76 d/y. This translates into an ice-cover reduction rate of 1.75 d/y, or about 4.5 times that found for the more southern

[1] In this context,'+' and '-' refer to later and earlier, respectively, and the use of different reference intervals for reporting changes in subsequent shorter-term studies is discussed in subsequent sections. It should also be noted that there are typographical errors in the table of record lengths for freeze-up reported by Magnuson et al. (2000), with the last digit of the 100+ records missing in the published article but contained in the Global Lake and River Ice Phenology Database (GLRIP).

parts of Canada for the most rapid depletion period of 1970 to 2004. The degree to which this reflects the effects of either more recent or higher-latitude warming or a combination of both is unclear. Explaining the reasons for any of the regional contrasts will require further investigation into factors such as regional differences in controlling heat fluxes, as well as differences that may arise from the use of different observational methods (e.g., *in situ* vs remote sensing).

In addition to basic observations of freeze-up and break-up timing, some trend analysis has also been conducted on other phenological characteristics of river-ice break-up. For example, in a study of rivers of the Mackenzie River Basin for the period 1970 to 2002, de Rham et al. (2008b) determined that both the initiation and timing of peak water levels occurred an average of 0.12 d/y earlier. Focusing specifically on the northern part of the basin, the Mackenzie River Delta, Goulding et al. (2009a) showed that spring streamflow pulse and the timing of melt initiation have advanced by 1.1 d/decade and 2.0 d/decade, respectively (both non-significant over the period 1974 to 2006). In addition, the ablation period (from melt initiation to the onset of break-up) has increased over the 33-year record. This agrees with results from Smith (2000) for Russian Arctic rivers, where a trend toward a longer pre-breakup was noted and suggested to be a potential driver of more frequent thermal break-ups in the future. Goulding et al. (2009a) also found a significant decrease in peak ice thickness of 4 cm/decade and a significant increase in freeze-up stage by 27 cm/decade, although considerable interannual variability was evident.

Large-scale, comprehensive records of river- and lake-ice thickness are rare. One dataset compiled for Canada over the past 50 years (Lenormand et al., 2002) does not reveal any obvious trends over the latter part of the 20th century (Lemke et al., 2007). However, reductions in maximum ice thickness of 2 to 14 cm have been observed on nearly all rivers and lakes within Arctic Russia over the past two decades compared to the previous 30-year period (Vuglinsky and Gronskaya, 2006). The largest decreases were found in rivers within Siberia (Ob: 5 to 10 cm; Yenisey: 6 to 14 cm; lower Lena: 11 to 15 cm). With reference to the interpretation of trends in ice-cover thickness, it should be pointed out that caution must be used when employing records from systems that have become regulated during the period of interest. For example, there has been a marked increase in ice thickness at the Tuoy-Haya station in Russia after the river site was impounded to become the Vilyuiskoeye reservoir (2360 km^2) (Figure 6.8). The increase can be attributed to changes in ice dynamics, hydraulics, and water-ice heat fluxes resulting from the storage. Including such records in any trend analysis of the effects of changing climate, without accounting for regulation-induced effects on ice growth, could lead to incorrect interpretations.

6.4.2.2. Linkage of trends to climatic factors

Freshwater ice duration, thickness, and composition are influenced by a diverse set of hydraulic and hydroclimatic variables that often span the entire ice season (see Section 6.3). Given this complexity, establishing relationships between freshwater ice trends and climate has generally been conducted simply using air temperature (see Section 6.2). Ice thickness is the variable most frequently linked to such a readily available, single climatic variable. For example, Vuglinsky and Gronskaya (2006) established a relationship between maximum river-ice thickness and winter air temperature in European Russia. Air temperature, however, has also been used to explain the more complex phenological timing of freeze-up and break-up. Statistically, air temperature during the preceding autumn and spring months is often able to explain 60% to 70% of the variance in the timing of ice break-up and freeze-up on lakes and rivers (Walsh et al., 2005). As noted by Prowse and Bonsal (2004), a first approximation of river-ice response to climatic change based on the analyses of various cold regions indicates that a long-term mean increase of 2 to 3 °C in autumn and spring air temperature has produced an approximate 10-day delay in freeze-up and a 15-day advance in break-up. For one specific period (1870 to 1950), it was found that a 3 °C rise in April temperature was associated with about a 15-day advance in break-up (Zachrisson, 1989). This concurs with the typical 5 d/°C rate of change in phenological data for many lakes and a few rivers around the Northern Hemisphere (Magnuson et al., 2000). As a result, trends toward shorter freshwater-ice duration over much of the circumpolar North closely correspond to the increasing air temperature trends observed over most of this region.

Another approach to assessing climatic linkages to freshwater-ice break-up and freeze-up timing has involved relating these events to 0 °C isotherm dates as defined by Bonsal and Prowse (2003). This variable has the advantage of not being constrained by the traditional seasonal definitions of temperature normally used in large-scale climatic studies. For several rivers within Canada, Lacroix et al. (2005) determined that break-up dates were highly correlated to the timing of spring 0 °C isotherms over most of the country; however, these relationships were much weaker and less spatially coherent during autumn. In terms of Canadian lakes, similar spatial and temporal patterns have been found between trends (1966 to 1995) in autumn and spring 0 °C isotherms and lake freeze-up and break-up dates. In general, this included significant trends toward earlier springs and earlier break-up dates over most of western Canada, and little change in the onset of lower temperatures and freeze-up dates over the majority of the country in autumn (Duguay et al., 2006).

Since air temperature varies sinusoidally over the year, the calendar dates on which the smoothed air temperature falls below 0 °C (in autumn) and rises above 0 °C (in spring) are arc cosine functions of the smoothed air temperature, and the fraction of the year during which the ambient air temperature is below 0 °C can be estimated by $(1/\pi)\ \arccos(T_m/T_a)$, where T_m and T_a are the mean and amplitude of the annual air temperature cycle (Weyhenmeyer et al., 2005; Livingstone et al., 2010). This arc cosine approach has been validated for more than 40 years of historical ice phenology data from 196 lakes spanning 13° of latitude in Sweden. Due to the form of this function, the timing of ice break-up (and freeze-up) tends to be more sensitive to variations in air temperature at lower latitudes where mean annual air temperatures are higher, than at higher latitudes where mean annual air temperatures are lower. In other words, the timing of lake-ice break-up appears to respond to changes in air temperature in a non-linear fashion dependent on latitude (Blenckner and Chen, 2003; Weyhenmeyer et al., 2004, 2005). As outlined in the previous section, this non-linear response has been identified in lake-ice trend analyses over Sweden and Finland. An opposite latitudinal trend might apply to the data in Canada, although the reasons for these opposing trends requires further investigation, as noted in Section 6.4.2.1.

6.4.2.3. Linkage of trends to atmospheric circulation patterns

Although freshwater ice-cover duration over the circumpolar North has, for the most part, significantly decreased in response to increasingly warmer conditions during the 20th century, the patterns have varied regionally, primarily due to climate impacts associated with large-scale atmospheric and oceanic oscillations. Teleconnection patterns originating over the Pacific and Atlantic Oceans account for a substantial amount of the observed 20th century northern-hemisphere temperature trends and variability, especially during winter and spring (Hurrell, 1996; Serreze et al., 2000). For example, the El Niño-La Niña / Southern Oscillation (ENSO), Pacific North American (PNA) pattern, and Pacific Decadal Oscillation (PDO) are closely linked to winter and early spring temperature variability over much of North America, particularly to the west. This includes higher temperatures associated with El Niño events and positive phases of the PNA and PDO, which are all representative of a deepened Aleutian Low (and vice versa) (e.g., Wallace et al., 1995; Bonsal et al., 2001). The most notable pattern in the Pacific involved a shift toward a deeper Aleutian Low after 1976 that has been associated with the trend toward higher winter and spring temperatures over western North America (e.g., Trenberth, 1990). The deeper low and associated cyclonic surface circulation allow for more frequent incursions of warm Pacific air into northwestern North America and thus shorter ice durations (especially, earlier break-ups) in this region (see Figure 6.9). In fact, several studies have demonstrated convincingly that ENSO, the PNA, and the PDO have had significant impacts on the ice phenology of Arctic lakes and rivers within North America including the mid-1970s interdecadal shift (e.g., Benson et al., 2000; Robertson et al., 2000; Bonsal et al., 2006; Blenckner et al., 2007).

In Europe, parts of northern Asia, and northeastern North America, the North Atlantic Oscillation (NAO) (or the Arctic Oscillation (AO) with which it is strongly associated) is known to influence large-scale climate, especially air temperature, in winter and spring (e.g., Hurrell, 1995). In particular, a positive NAO

(representative of a deepened Icelandic Low in the North Atlantic) is associated with higher temperatures over Europe and northern Asia and lower than normal values over northeastern Canada (and vice versa) (Hurrell, 1996). The intensified counter-clockwise circulation of the Icelandic Low results in cold northerly air advection into eastern Canada and greater frequencies of warm maritime flow into Europe and northern Asia (Figure 6.9). As a result, a strong winter NAO/AO signal has been detected in the ice phenology of lakes and rivers in Estonia, Finland, Sweden, and northwestern Russia (Livingstone, 2000; Yoo and D'Odorico, 2002; Blenckner et al., 2004, 2007; Karetnikov and Naumenko, 2008), and for Lake Baikal in Siberia (Livingstone, 1999; Todd and Mackay, 2003). These oscillations have also been shown to impact freshwater-ice duration in northeastern Canada (Bonsal et al., 2006). A higher occurrence of positive NAO and AO in the later decades of the 20th century is consistent with the winter and spring cooling observed over most of northeastern North America and warming in Europe and northern Asia. This, along with the Aleutian Low shift in 1976, helps explain the west to east gradient in air temperature and associated freshwater-ice duration observed over Arctic regions of North America during the latter 50 years of the 20th century. It is also consistent with the observed trends toward shorter ice duration in much of Scandinavia and western Russia. Notably, from around 2000 to the present, persistent warm temperature anomalies have occurred in two localized areas, including eastern Siberia and northeastern Canada and Baffin Bay, which is in contrast to temperatures observed during the latter part of the 20th century. This is the result of a northward displacement and strengthening of the Aleutian Low and a weakening of the Icelandic Low, the latter being reflective of a trend toward more negative NAO and AO values since the turn of the century (Overland and Wang, 2005). Although not specifically analyzed, it is likely that these two regions have been associated with shorter freshwater-ice duration during this period.

The connection between the Pacific and Atlantic large-scale oscillations shown in Figure 6.9 were not consistent through the 20th century, although some recent studies have indicated a seesaw relationship linking the Aleutian and Icelandic Lows that acts on interdecadal timescales (e.g., Honda and Nakamura, 2001; Luchin et al., 2002). The linkage was weak from the mid-1950s through the mid-1960s and particularly strong from the 1920s to the 1940s and from the late 1960s through the 1990s. It consists of Aleutian Low anomalies of a given sign developing in mid-winter followed by the occurrence (through Rossby wave propagation) of Icelandic Low anomalies with opposite sign about one month later (Honda and Nakamura, 2001). This then affects surface temperatures over northwestern North America, Europe, and northern Asia (Honda et al., 2005). The seesaw in surface pressure and resultant impacts on surface temperature (see Figure 6.9) are likely to be reflected in results from Pavelsky and Smith (2004) who determined that break-up dates on the Mackenzie and Lena Rivers (from 1992 to 2003) showed a consistent negative correlation. They attributed this relationship to variability in the PDO, where in seven of the ten years a positive PDO index (i.e., deeper Aleutian Low) was associated with earlier break-up on the Mackenzie and later ice-off dates on the Lena (consistent with a weaker than normal Icelandic Low). Further research is required to determine co-relationships between freshwater-ice dates over various regions of the circumpolar North and corresponding linkage to large-scale atmospheric circulation patterns.

It should be noted that recent extreme warm periods over the Arctic have been attributed to distinct atmospheric circulation patterns that are generally not captured by the standard teleconnection indices. The best-documented case has been the summer of 2007 when anomalously high temperatures and associated extreme sea-ice loss over the western Arctic were linked both to a very unusual atmospheric circulation pattern that resembled the PNA, but was much stronger and shifted northward (L'Heureux et al., 2008), and to a dipolar circulation pattern consisting of a northeastward shift to the AO and NAO centers of action, dubbed the Arctic Rapid Change Pattern (ARP) by Zhang et al. (2008). These recent atmospheric 'hot spots' may be suggestive of a new climate state for the Arctic (Overland et al., 2008), thus making future changes to freshwater-ice duration more uncertain.

In addition to large-scale teleconnections, a few studies have attempted to relate freshwater ice phenology to regional or synoptic atmospheric circulation patterns. For example, in a study of 50 lakes in Sweden and Finland ranging from 58° to 69° N for the period 1961 to 2002, Blenckner et al. (2004) found that

regional zonal and meridional circulation indices explained a larger portion of observed variability in freeze-up and especially break-up dates when compared to the larger-scale autumn and winter NAO index. It was therefore concluded that regional circulation indices are a useful tool for climate impact assessment of freshwater ice and require additional examination over other regions within the circumpolar North.

6.4.3. Projected changes

- Recently, there have been concerted efforts to model changes to future freshwater ice phenologies using climate output from general circulation models (GCMs) and regional climate models (RCMs). For example, modeling of hypothetical lakes of varying depth for northern land masses between 40° and 75° N indicates that future warming by the period 2040 to 2079 will result in the following:
  - an overall increase in lake water temperature, with summer stratification starting earlier and extending later into the year, resulting in the timing of freeze-up being delayed by 5 to 20 days
  - break-ups that are 10 to 30 days earlier
  - a decrease in lake-ice duration of about 15 to 50 days
  - decreases in maximum lake-ice thickness by 10 to 50 cm
  - a likely increase in surface white ice, especially as the ice cover also thins (owing to projected increases in winter snowfall at high latitudes).

- Modeling of future changes to river-ice regimes has been restricted to a few rivers, with no regional studies conducted.

- Projected decreases in south-north gradients in air temperature suggest that the severity of break-up and related ice-jam flooding may be reduced on some large, northward-flowing rivers, but effects might be mitigated by other factors, including changes in the magnitude of spring snowmelt.

Similar to the assessment of historical trends in freshwater-ice characteristics (Section 6.4.2), the quantification of future changes to lake and river ice is primarily based on degree-day methodologies as inferred by projected temperature changes from climate models. Results from these studies show a continued reduction in freshwater-ice duration (of varying degrees) in all Northern regions. For example, over European Russia and western Siberia, Borshch et al. (2001) applied a uniform warming of 2 °C to assess changes in river-ice duration. It was determined that break-up would be advanced by 4 to 10 days and freeze-up delayed by 4 to 12 days, with the greatest changes in western regions of Russia. Relying on the 5 d/°C rate of change in phenological break-up dates estimated by Magnuson et al. (2000), Prowse et al. (2002a) approximated that a projected increase in spring air temperature of 3 to 7 °C by the end of this century (as determined from several GCMs) would result in a 15- to 35-day advance in river-ice break-up in northern regions of Canada. A more detailed analysis based on average monthly temperature change projections (relative to 1961–1990 using the IPCC A2 emissions scenario from seven GCMs) for the 30-year period centered on the 2050s (2040–2069), found that spring 0 °C isotherm dates will occur 6 to 10 days earlier over northern regions of Canada. Autumn changes were even more pronounced (~10 to 12 days later) (Prowse et al., 2007a; Figure 6.10). Given the close correspondence between 0 °C isotherm dates and lake-ice (e.g., Duguay et al., 2006) and river-ice (e.g., Lacroix et al., 2005) break-up and freeze-up dates during the instrumental record, it may be hypothesized that by the middle of the 21st century freshwater-ice duration over much of northern Canada will be about 20 days shorter than during the 1961 to 1990 baseline period (Prowse et al., 2007a). Subsequent analyses of future 0 °C isotherm changes along the four major Arctic rivers (Lena, Mackenzie, Ob, Yenisey) by Prowse et al. (2010) indicate that pronounced changes in river ice will occur along some of these rivers.

As noted by Bonsal and Prowse (2003), however, these empirical relationships between freshwater-ice dates and air-temperature indices may not be reliable under future climatic conditions due to changes in the composition of major heat fluxes on which the temperature relationships are founded. Furthermore, as outlined in Section 6.4.2, the historical timing of freshwater-ice break-up in some regions of the North has responded to changes in air temperature in a non-linear fashion that was dependent on latitude (Blenckner

and Chen, 2003; Weyhenmeyer et al., 2004, 2005), thus complicating the use of spatially uniform degree-day and freshwater-ice change relationships.

There have only been a few preliminary analyses that have modeled changes in future freshwater-ice phenologies using climate output from GCMs. In the case of lakes, Dibike et al. (2010) incorporated a one-dimensional lake simulation model (MyLake) and atmospheric forcing data from the ERA-40 global re-analysis dataset (Uppala et al., 2005) to simulate lake-ice phenology and composition for all land masses in a 40° to 75° latitudinal band with hypothetical lakes positioned at a resolution of 2.5° latitude and longitude. Differences in driving climatic variables between a current (1960–1999) and future (2040–2079) run of the Canadian Global Climate Model (CGCM3) were applied to the ERA-40 current data as input to the MyLake model to project future ice conditions. Results indicated that future warming will result in an overall increase in lake-water temperature, with summer stratification starting earlier and extending later into the year and, hence, the timing of freeze-up being delayed by 5 to 20 days (Figure 6.11). Break-up was projected to occur 10 to 30 days earlier, resulting in an overall decrease in lake-ice duration of about 15 to 50 days. Maximum lake-ice thickness was also modeled to decrease by 10 to 50 cm. Change in snow loads and related ice-cover composition were also modeled. In general, maximum snow depth changed by -20 to +10 cm and white ice by -20 to +5 cm, depending on the geographical location and other climate parameters – the high latitudes being an area of projected increases in winter snowfall that can promote white-ice formation particularly with thinner ice cover.

There are fewer projections of future climate change impacts on river ice than lake ice, and none have been conducted at regional scales. One large-river example is provided by Andrishak and Hicks (2008), who applied a one-dimensional hydrodynamic model to assess climate change impacts on ice-cover extent and duration on the Peace River in Canada. Incorporation of air-temperature projections (for the 2050s using the IPCC A2 emissions scenario) from the second generation Canadian Global Climate Model resulted in an average reduction in ice duration of 28 days (13 days later for freeze-up and 15 days earlier for break-up).

Much less is known about changes in more complex variables such as ice composition and, in the case of rivers, the frequency and severity of ice jams. One of the more important potential changes to river ice relates to the severity of break-up advance (see also Section 6.5.1.1). Whether temporal shifts in river-ice duration will produce more or less severe break-up events (i.e., floods) remains unknown, largely because of the complicating role of precipitation, which has the potential both to control the driving (snowmelt runoff) and resisting (ice thickness, strength, composition) forces that affect break-up severity. Although some very limited, site-specific predictions of change in these two cryospheric components (snow and river ice) have been made to estimate the effect on river-ice break-up severity (Beltaos et al., 2006), broad-scale analyses have not been undertaken.

As suggested by Prowse et al. (2006), changes in the thermal gradients affecting the advance of the spring freshet on northward-flowing rivers could produce significant changes to the dynamics and related flooding that accompany river-ice break-up. Recent analysis of 0 °C isotherms along the four major Arctic rivers (Lena, Mackenzie, Ob, Yenisey) by Prowse et al. (2010) suggests that future warming could lead to a reduction in the thermal gradients. The 0 °C isotherm was used because it was noted to approximate the timing of river-ice break-up (e.g., Bonsal and Prowse, 2003; Lacroix et al., 2005; see also Section 6.4.2.2). Figure 6.12 illustrates average projected changes in 0 °C conditions on the four largest Arctic rivers based on average values of four GCMs for two future time periods, 2041–2070 and 2071–2100, and referenced to current climate (1979–2008). All four rivers exhibit progressively earlier timing of the 0 °C isotherm over their entire length. They also show a tendency to greater warming in a downstream (primarily south to north) direction, with the greatest change being for the Lena River. Such a reduction in the current climatic gradient is likely to lead to a more thermal type of break-up (see Section 6.5.1.1 for a description of break-up dynamics) characterized by reduced ice action and ice-jam flooding. This would have major ecological implications for riparian ecosystems, such as river deltas (see Section 6.5.2.3).

Given that large-scale teleconnections have been shown to influence air temperature and associated river-ice characteristics over much of the Northern Hemisphere (Section 6.4.2.3), knowledge of future changes in the frequency and magnitude of these oscillations could provide insight into future freshwater-ice regimes. At present, however, the effects of climate change on large-scale teleconnection patterns remain uncertain due to the lack of agreement concerning the future frequency and structure of atmospheric and oceanic modes among the various climate models (Prowse et al., 2007a). For example, with respect to El Niño, GCMs from the most recent (2007) IPCC assessment offer a wide range of possibilities with respect to the future occurrence and variability of these events (Guilyardi et al., 2009). Because El Niño significantly affects climate over many regions of the Northern Hemisphere and is often associated with variations in other Pacific-related teleconnections, such as the PDO and PNA, knowledge of its future occurrence would assist in projecting regional changes in future river-ice duration and dynamics. The majority of climate change investigations have indicated increased occurrence in the positive phases of the NAO and AO in association with future warming (e.g., Rind et al., 2005). This would result in continued warming over northern Eurasia and cooling over northeastern North America with implications for freshwater-ice duration over these regions (see Figure 6.9).

In summary, predicting the state and fate of freshwater ice over the next century will require a number of significant advances, the most difficult being for river ice. For the simpler case of lake ice, improvements will need to be made in the physical modeling of the full-season thermal regime. This is especially important in determining the timing of freeze-up and in considering the increasing role of precipitation (e.g., snow accumulation) that affects ice-growth rates, ice-cover composition, ablation rates, and the timing of break-up. As for most other surface variables, refinements will be made to properly downscale (statistically or dynamically) the requisite variables that control surface energy exchanges, including atmospheric coupling to account for feedback in the case of large lakes. Site-specific to broad-scale predictions of change in most lake-ice characteristics would be possible with the advancements noted above (Prowse et al., 2008).

6.5. Effects of changing freshwater ice covers

6.5.1. Hydrological effects

- Many projected changes to freshwater ice systems will be affected by companion changes in other cryospheric components, particularly snow and permafrost.

- Changes in break-up severity are a major concern because of their importance to flooding on northern rivers, which can have positive and negative impacts.

- The location and severity of river-ice break-up could also be modified by changes in hydraulic gradients, particularly in coastal deltas affected by rising sea level.

- Mid-winter break-ups will increasingly intrude into higher latitudes; there is already evidence of such events occurring in the sub-Arctic.

- Ice-regime changes are also likely to affect river low-flows through modifications of ice-induced hydraulic storage and icing accumulation / ablation that have historically supplied flow during low-flow summer months.

- Decreases in ice duration combined with higher summer temperatures will lead to increased lake evaporation, lowering of lake levels, and potential drying out of shallow basins. Recent evidence suggests that this may have already happened to some ponds that have been permanent water bodies for millennia.

- Winter pulsing of water from lakes to downstream rivers will occur as ice-cover thins and winter precipitation loads increase.

A brief physical description of river ice processes was provided in Section 6.1.2, but to aid in understanding how climate change may affect river-ice flow regimes the following subsections provide more details about how key climatic variables control hydraulic and mechanical conditions associated with some specific ice-induced hydrological extremes. There are, however, a number of other cryo-hydrological changes that will affect freshwater-ice systems. To varying degrees, such changes have been reviewed in other chapters, particularly those dealing with snow and permafrost. The major changes are briefly summarized in this introductory section and considered in more detail in subsequent sections.

One of the most significant changes in other cryospheric components that will affect both lake and river ice is the change in winter precipitation. Recent research (e.g., Zhang et al., 2007) suggests that there has been a global latitudinal redistribution of precipitation from lower to higher latitudes. Moreover, future projections for higher latitudes suggest that such changes will mean increases in winter precipitation and snow-water equivalents (Bates et al., 2008; Brown and Mote, 2009). This will also result in greater snow loading of freshwater-ice covers and, as explained below, could affect thermal insulation influencing ice-growth rates, ice-cover depression and the formation of surface snow ice, and winter pulsing of flow displaced from lakes. Increased winter snow accumulation, unless interrupted by mid-winter melt, should also translate into a larger spring snowpack. As already documented for some large northern rivers and alpine catchments, both the magnitude and timing of spring discharge are strongly linked to change in snow cover during the spring melt season (e.g., Yang et al., 2007, 2009; Stewart, 2009). Assuming that future spring-melt conditions (e.g., intensity and duration of heat fluxes) are similar to those for the current climate, a larger snowpack should translate into a larger melt volume and river discharge. Several studies have shown that winter discharge in the major Eurasian Arctic rivers is already increasing (see White et al., 2007). Larger winter baseflow and spring snowmelt volumes could increase the forces that drive river ice and have major implications for break-up severity, javes (a steep water wave associated with the release of an ice jam on a river), and ice-jam floods. If sufficiently large, mid-winter melt of a catchment snowpack also has the potential to initiate ice break-up.

While changes in snowfall are most important to spring ice events, other seasonal changes in river flow created by changing permafrost conditions (e.g., Hinzman et al., 2005; White et al., 2007) can also have implications for freshwater-ice processes. Overall, thawing permafrost and the associated development of more groundwater flow systems are projected to increase river baseflow, some of which has already occurred during winter in parts of the Northwest Territories, Canada (St. Jacques and Sauchyn, 2009). Combined with any summer and autumn increases in precipitation, this should lead to higher flows at the time of winter ice formation. Higher flows and stage at freeze-up raise the possibility of thicker ice accumulations, increased hydraulic storage of water, and even the potential for autumn freeze-up ice jams and related flooding (Beltaos and Prowse, 2009). Increased groundwater flow will also mean more heat transfer to rivers, which should translate into some thinning of ice cover, or where flow is concentrated, even the opening or broadening of freshwater polynyas (e.g., Prowse, 2001b). Increased groundwater flow will also thin ice cover in lakes, creating more unfrozen volume beneath the ice. This has important implications, especially for shallow systems that freeze to the bed, for the availability of under-ice winter habitat, and for the migration of fish species. In extreme cases of permafrost thaw and groundwater development, talik development (i.e., zones of localized unfrozen ground) beneath lakes could lead to their drainage and the elimination of ice covers.

Many hydrological changes operating in the Arctic are controlled by synergies among cryospheric components. Further details of many of these are reviewed in the following sections.

6.5.1.1. River-ice floods and javes

Large aggregate thickness and extreme underside roughness are two characteristics of river-ice jams that combine to generate very high water levels relative to those occurring for the same discharge under open-water conditions. Although freeze-up jams can cause flooding under certain hydro-climatic conditions, it is break-up jamming that typically generates the most extreme floods, owing to the much higher break-up flows (Beltaos, 2008d). The release of a major jam is also of concern since it is often attended by a steep water wave that can be metres high and is characterized by greatly amplified flow velocities and hydrodynamic forces. This release is also known as a jave, short for 'jam release wave' (Beltaos, 2008e). Ice jams and javes have many socio-economic and ecological impacts, not all of which are negative (Sections 6.1.4, 6.5.2 and 6.5.3). For example, water- and ice-related damage as well as disruption of aquatic life and habitat are partially balanced by the replenishment of fragile ecosystems with water and nutrients. River ice jams are known to occur in all Arctic and sub-Arctic regions (Catalogue on ice jams in the former USSR, 1978; Beltaos et al., 1990; White and Eames, 1999; Beltaos and Prowse, 2009).

As soon as mild weather sets in (normally in spring), river-ice covers begin to experience thermal and hydro-mechanical change. Factors that *resist* ice-cover dislodgement, such as thickness, strength and attachment to channel boundaries, diminish. At the same time, factors that promote or *drive* dislodgement, such as hydrodynamic forces and water-surface width, increase. This concept is illustrated in Figure 6.13, which traces the evolution of driving and resisting factors, termed 'forces' for simplicity, during the pre-breakup period. Break-up is initiated when the driving force exceeds the resisting force. If the resisting force (point A) is near the lower limit of the driving force, negligible or no jamming occurs, and the break-up event is termed 'thermal' to signify that it is dominated by thermally induced decay of the ice cover. Non-thermal events are called 'mechanical' because the ice cover retains a significant amount of its mechanical competence when it is first dislodged. The upper part of the mechanical range is termed 'dynamic' because the break-up is now dominated by the large driving forces. Major ice jams and javes typically occur during dynamic break-ups.

In the Arctic, dynamic break-up and jamming events are typically driven by large spring runoff that primarily results from rapid snowmelt, which often occurs in southern, sub-Arctic portions of large river basins. Where the general flow direction is northward, the rising flows encounter increasingly competent ice cover, owing to spatial gradients in air temperature. Rainfall may be a significant spring runoff factor in sub-Arctic regions, and a dominant one for mid-winter break-ups. The latter are triggered by rain-on-snow events and lead to highly dynamic conditions.

Changes in the flooding potential of ice jams and javes are difficult to detect and quantify because they involve highly complex phenomena and hydro-climatic interactions. As a result, relevant literature about climate interactions and the severity of ice jams, particularly at high latitudes, is relatively scarce. The following is indicative of the current knowledge base. In northwestern Canada, a slight trend toward increasing peak annual break-up water levels beginning in the 1970s has been detected for the Yukon River at Dawson (latitude ~64° N) (Janowicz, 2009). Ice break-up on Alaskan rivers moderated during the 1995 to 2005 period, but the particularly destructive 2009 break-up (Janowicz, 2009) may be heralding a different trend. In sub-Arctic Canada, trends toward more frequent mid-winter floods and more severe spring floods have been detected in a few case studies (Beltaos, 2002, 2004; Prowse et al., 2002a).

On Russian rivers, the risk of ice-jam flooding increases from south to north and west to east, while there is a temporal trend toward more severe events on rivers in eastern Siberia (Buzin, 2007, 2008). This is exemplified by the Lena River (Figure 6.14), where the catastrophic flood of 2001 devastated the city of Lensk. Elsewhere in Russia, there is a diverse pattern of change. In some regions (northern and central European territories of Russia and western Siberia), almost no change has been observed. For Scandinavia, mid-winter break-up events in Norway (Knut Alfredsen, [affiliation], pers. comm., 2009) appear to be occurring more frequently and with more severity in recent years, but similar to most other regions, no analysis has yet been published.

It is difficult to project how a changing climate will affect the frequency and severity of ice-jam floods and javes, owing to the multitude of relevant hydro-climatic controls (Section 6.3.2), although some attempts have been made. Beltaos and Prowse (2009) discussed how each climatic control may affect ice-jam regimes under a changing climate, but cautioned that projection of the synergistic effects of all controls requires detailed, site-specific studies. For example, the projected changes in thermal gradients along rivers are likely to affect river-ice dynamics (see Section 6.4.3). This effect may be negated, however, by an earlier melt, which will result in reduced pre-breakup insolation and associated ice decay. The latter effect is illustrated by the dashed lines in Figure 6.13, which indicate that (other factors being equal) the pre-breakup decrease in resisting force would be less pronounced under an earlier-melt regime.

A relatively general projection is that mid-winter break-up and associated dynamic ice jams will increasingly intrude into higher latitudes of the Northern Hemisphere (Prowse et al., 2002a; Beltaos and Burrell, 2003; Prowse and Bonsal, 2004; Beltaos and Prowse, 2009). Such intrusion is not likely to affect Arctic areas where the winters will remain uniformly cold despite future warming, but there is already evidence of the northward shift of mid-winter break-up. On the Klondike River (Yukon Territory, Canada), an unusual period of mild weather and rain in December 2002 resulted in early break-up and jamming (Janowicz, 2009). Although this jam only caused minor flooding at the time, it froze in place when the cold weather resumed, creating a thick 'plug' with major spring break-up implications: in late April 2003, the lower Klondike valley experienced one of the most severe break-up floods on record.

For Norway, downscaled scenarios from climate models suggest that more frequent ice runs are likely to occur in the future, with possible jamming at new locations (Asvall, 2009), whereas more 'frazil-risk' days can be expected in Finland (Huokuna et al., 2009). Some attempts have been made to predict changes in future ice-jam flood conditions for Russian rivers. For example, research at the Russian State Hydrological Institute evaluated conditions for the very near-term period of 2010 to 2015, based on the results of the HadCM3 and ECHAM4 GCMs for two climate change scenarios. Changes to flood magnitudes were assessed using regional correlations with river discharge. The following describes the major results regarding ice jams, as summarized by Buzin (2007, 2008) and Buzin and Kopaliani (2007). In general, the authors concluded that with the expected rise in river discharge in winter, there would be an increased probability of major ice jams in spring. Some of this is due to change in streamwise thermal gradients (see Section 6.4.3). A recently discovered feature of downstream ice clearance is that the break-up front on the northward-flowing Russian rivers is often delayed at latitudes of 58° to 60° N due to sharp spatial gradients in spring air temperature. At this point under an altered future climate, the positive-sign anomalies in the upper reaches of the rivers change to negative-sign anomalies further downstream and result in powerful ice jams. The projection shows that spring ice-jam floods on the Lena River and other Siberian rivers will become more frequent (1.2 to 1.5 times more frequent) and more severe (a 35% to 60% increase in peak break-up water levels). In northern parts of the European territories of Russia, recurrent ice jams are expected on the rivers Severnaya Dvina, Sukhona, Vug, and Pechora. On average, maximum jam levels might increase 10% to 12% and on individual river sections 36% to 48%. However, the frequency of flooding is projected to increase by a factor of 1.2, at most. Many populated areas on the shores of these rivers (Shenkursk, Kholmogory, Archangelsk, Naryan-Mar) are expected to be periodically flooded. In this region, there is also a threat of powerful spring ice jams on the Sukhona River near the city of Veliky Ustyug. The water level elevation caused by an ice jam would increase 24% to 36%, compared to the highest level under stationary conditions. These estimates of potential change in the frequency and scale of ice-jam floods should be considered as averaged over large territories, and may be different for individual rivers and river reaches.

The location and severity of ice jamming can be modified not only by changes in thermal gradients (see Section 6.4.3), but also by changes in hydraulic gradients. While differential changes in water levels among intersecting rivers and lakes could have this effect, the largest changes in hydraulic gradient are likely to result at river mouths and deltas that enter marine systems (e.g., see Arctic mega delta discussion by Anisimov et al., 2007). This is because of the anticipated rise in sea level, which is projected to be significant even over the next 100 years. Increases as high as those predicted by the IPCC (e.g., 0.18 to

0.59 m; Bates et al., 2008) or in more recent analyses that place more focus on rapid changes in ice-sheet flow (e.g., 0.7 to 2.0 m; Pfeffer et al., 2008; Grinsted et al., 2010; also see the Integration Chapter on Sea Level Rise) are more than ample for the sea level to extend significantly back up into low-slope, Arctic river deltas. The net rise (relative to the land) may be moderated or enhanced by concurrent vertical motions of the land surface. For example, glacio-isostatic adjustment varies considerably around the Arctic, from uplift in the Canadian Archipelago, Greenland and Norway, to subsidence along the Beaufort Sea and Siberian coasts (Walsh et al., 2005). Tectonic motion and local loading of the crust by sediment (e.g., in river deltas) can also cause vertical adjustments.

Overall, a net rise in sea level would shift the location where the river water surface meets the largely flat profile of the sea-controlled channel (Beltaos and Prowse, 2001). At such sites, the water surface slope decreases sharply, thereby enhancing hydraulic conditions conducive to ice jam formation (e.g., Beltaos, 1995b). The higher the base level rise, the further upstream such effects are likely to migrate (Figure 6.15). Riverside communities and nearby infrastructure may be particularly vulnerable. At the same time, riparian ecosystems that depend on regular flooding generated by ice jams for replenishment could also be affected (see Section 6.5.2.3). Such changes could be further exacerbated by a related upstream increase of naturally-built levee height from lateral, ice-jam induced sediment deposition (Hill et al., 2001), which would in turn lead to reduced frequency of overbank flooding in areas still subject to ice jamming.

In general, projections for specific rivers and sites will require quantitative analysis of many factors, in addition to purely climatic variables that can be furnished by GCMs and RCMs under specific greenhouse gas emissions scenarios. Such factors include the flow hydrograph during the entire ice season (autumn, winter, spring), freeze-up levels, ice-cover thickness and strength, as well as river planform, bathymetry, and hydraulic characteristics. The effects of changes in these factors on the severity and frequency of ice-jam floods could be assessed using numerical process models (Liu et al., 2006). However, serious gaps in the quantitative knowledge of break-up processes remain an impediment to accurate prediction (Beltaos, 2008e).

6.5.1.2. River-flow abstraction

Hydrological extremes produced by ice on northern rivers include not only flood-related phenomena, but also low flows. Flow minima are especially important where rivers are used, for example, to supply municipal or industrial users or in the dilution of wastewater. The occurrence and magnitude of low flows during the ice season are directly affected by ice formation, which entails three types of flow abstraction via temporary storage of water (Gerard, 1981, 1990; Beltaos and Prowse, 2009), all of which are sensitive to climatic variability and change. The first and most obvious is the storage due to freezing of river water to form the ice cover. This effect is most pronounced shortly after freeze-up when ice growth is most rapid. Although generally negligible on large rivers, this type of abstraction can be significant in very small streams, especially during the freeze-up period. Less obvious, but far more pronounced, are the two other abstraction mechanisms: icings and hydraulic storage.

Icings (Section 6.1.2) can be so extensive in some northern systems that they significantly reduce the magnitude of flow on downstream larger-order rivers (e.g., Gerard, 1981; Sokolov, 1986). Icings effectively sequester baseflow during the winter months, thereby enhancing low flows, and then release water during the summer. Yoshikawa et al. (2007) estimated that an icing field in the Kuparuk River in Alaska stores around 27% of potential winter baseflow. Some accumulations are so large that they control the routing of flow during spring melt and sustain flow during the warmer summer months (e.g., Grey and MacKay, 1979). In extreme cases, icings tend to be relatively permanent features of rivers. For example, the Kongukut River in Alaska has an existing icing field that was first observed by the Franklin party in 1828. Aerial photography shows that this and many other icings in Alaska have not changed in extent during the past 50 years (Yoshikawa et al., 2007), although their future existence and role in the hydrological cycle depends on future climatic conditions.

The hydraulic storage of water also greatly influences low-flow conditions both on large and small Arctic rivers. In general, the presence of an ice cover causes channel water levels to rise, owing to its additional hydraulic resistance and flotation depth (keel). As the ice cover continues to propagate upstream, a fraction of the incoming flow is abstracted to fill the space created by the higher stage and, consequently, downstream flow is reduced. Beltaos (2009) showed that relative abstraction (expressed as a fraction of the incoming flow) increases with increasing concentration of ice arriving at the upstream edge of the ice cover, but decreases with increasing thickness of the newly formed cover. Hydraulic storage abstraction ceases once the process of ice-cover formation is completed although this can last for many weeks during the autumn freeze-up period on large northern rivers. Importantly, this ice-induced low flow can be the flow minimum for the year, even lower than that which occurs during the main winter period when landscape runoff is at a minimum (Prowse and Carter, 2002). Moreover, the final release of this water at the time of the succeeding break-up event can significantly augment the spring snowmelt event, although this is rarely accounted for in spring freshet analyses. On the Mackenzie River in Canada, for example, Prowse and Carter (2002) found that almost 20% of the spring freshet could be due to the release of water abstracted during the previous autumn freeze-up (Figure 6.16) and the percentage would be even higher if the melt of ice-stored water was also included.

Icings and freeze-up are both highly sensitive to change in heat exchange at the river surface, and hence the amount of ice forming. Other factors being equal, warmer freeze-up weather would result in reduced abstraction and, thus, improved low-flow conditions. If icings shrink or even disappear, however, then the seasonal hydrograph on icing-dominated systems will change, much in the same way as when glaciers disappear and are no longer available to supply baseflow during the late summer dry periods (see Chapter 7). This could have important implications for aquatic biota (see Section 6.5.2.2), especially if the icings (e.g., those largely accumulating from groundwater sources) also play an important role in determining stream water chemistry.

An indirect climatic control pertains to freeze-up discharge, which influences the initial thickness of the ice cover. Although projections remain to be finalized (Bates et al., 2008), it is generally believed that because of changes in precipitation and permafrost regimes, autumn flows are likely to increase in northern river systems. Higher flow velocities are likely to lead to thicker ice accumulations, but given that ice thickness is also likely to decrease because of generally warmer temperatures, the final combined effect of these two factors on ice-induced low flows remains uncertain.

6.5.1.3. Lake water budgets and flow enhancement

Lake ice plays an important role in the evaporation regime of northern lakes, as was reviewed in Section 6.3. Specifically, its reduction with climatic warming, along with increased summer heating of the open water, should increase lake evaporation. The significance of this process is likely to be more important for shallower lakes than deeper lakes, at least in the early stages of climatic warming. Overall, shallower lakes are more sensitive to enhanced warming because of the shorter duration of their ice cover and more rapid heating during the open-water period due to their lower volume to surface area ratios. Greater total heating will mean proportionately more evaporation and, when combined with their shallower depths, translate into larger relative decreases in water level. Prowse et al. (2006) suggested that some very shallow (e.g., <1 to 2 m) northern basins are likely to dry out and possibly become athalassic saline systems, unless there is a compensating increase in water inflow from increased precipitation. This may already be occurring based on evidence provided by Smol and Douglas (2007b). They suggest that, despite wetter conditions, a recent increase in evaporation due to higher temperatures and extended ice-free conditions has led to the complete summer drying of some ponds in the Canadian High Arctic that had been permanent water bodies for millennia. By contrast, for winter periods decreases in ice-cover thickness are likely to increase the unfrozen water volume of some lake systems, particularly those that currently freeze to the bed (e.g., Figure 6.17).

Change in precipitation on lake-ice covers is also apt to produce an additional hydrological effect via snow loading and volume displacement. Winter loading of lake ice by snow tends to depress the ice cover and force water from the lake basin into outlet streams, causing an increase in downstream discharge (e.g., Prowse, 2005, 2009). Such additions are important modifiers of the winter flow regime, particularly in areas of low winter discharge. The effect increases in direct proportion to the snow load to ice thickness ratio. Assuming that future climate will lead to an increase in snow load and a decrease in ice thickness, winter flows from northern lakes should be enhanced, although the timing and magnitude will be determined by the characteristics of future winter precipitation events. For example, if future precipitation is concentrated into fewer higher magnitude events, as some models predict, winter spikes might develop in the flow of northern rivers.

6.5.1.4. Sediment transport and geomorphology

Although lake ice is known to produce some geomorphological effects on lake shorelines, the dynamics of river ice are most important to the fluxes of sediment and changes in riverine morphology (e.g., Prowse, 2005). Climate change is likely to cause major changes in river ice dynamics (see Sections 6.4.3 and 6.5.1.1), particularly along the long northward-flowing rivers. The Arctic Climate Impact Assessment (Walsh et al., 2005; Wrona et al., 2005) noted that changes in river ice regimes would affect channel-forming processes, suspended sediment transport to the Arctic Ocean, and the productivity of riparian, delta, estuarine, and marine shelves near coastal margins.

As context for understanding how the geomorphological role of river ice will change under an altered climate, it is important to recognize that ice plays a dual role in high-latitude fluvial geomorphology, acting as a resisting force keeping material intact and as a driving or erosive force. An example of the former is where ice is frozen to the stream bed (e.g., anchor ice) and effectively shelters sediments from the erosive actions of streamflow (e.g., McNamara et al., 2008). The loss of protective layers of bedfast ice under a warming climate could, therefore, produce enhanced erosion.

Significant changes are also likely to result from increases or decreases in the erosive forces of ice, particularly those associated with javes and ice jamming, which are characterized by high stage and flow velocities (Section 6.5.1.1). As reviewed by Prowse (2005), some of the major effects of ice as a driving force include the following:

- concentration of flow and bed scour leading to changes in the position and/or depth of the thalweg (deepest line of the river channel)
- suspended sediment concentrations during break-up that can be several times the equivalent discharge under open-water conditions
- increased material size and magnitude of bed load transport from break-up javes
- creation of erosional features such as high-level benches and undercut banks
- over-steepening of banks and initiation of slope failures, particularly in areas where ice-rich permafrost is exposed
- depositional features along banks varying from localized boulder accumulations (boulder buttresses, barricades, ridges, pavements) to thick layers of sediment, which can raise bank heights and are a particular development feature of northern deltas.

One overarching and highly important geomorphical effect of river ice is its erosive action in modifying channel widths. While some studies have hypothesized that channels are enlarged by break-up ice erosion accompanying infrequent high-stage events, other evidence has suggested that overbank losses of flow due to ice jams may even promote channel narrowing (reviewed by Prowse, 2005). More recently, however, additional information (Boucher et al., 2009; McNamara and Kane, 2009) confirms that ice erosion can produce enlarged channels if a minimum ice-jam frequency is exceeded. Hence, it is likely that changes in the break-up dynamics of northward-flowing rivers will have major consequences not only for their sediment transport regimes, but also for their overall morphology. Morphological features that

affect the capacity of a channel to convey floodwater and those that affect habitat are most significant (Prowse and Culp, 2008).

6.5.2. Ecological effects

- Changes in the timing of freeze-up and break-up on lakes will affect a wide range of related biological aspects of seasonality.
- Some changes are likely to be gradual, but others are likely to be more abrupt as systems cross critical ecological thresholds; positive and negative effects will result.
- Changes in ice-induced hydrological connectivity and lake stratification could lead to the loss of some species and the establishment of others.
- Changes in river ice are also likely to have wide-ranging effects on the behavior and biological response of stream biota; positive and negative effects will result.
- Changes in river dynamics associated with break-up are of particular concern because of their effect on river geomorphology, vegetation, sediment and nutrient fluxes, and sustainment of riparian aquatic habitats.

6.5.2.1. Lentic systems

High-latitude lakes are covered by ice for six to twelve months each year, and this has a wide-ranging influence on their ecology. Future changes in lake-ice conditions may cause shifts in physical, chemical, and biological properties, including increases in the duration of open-water conditions and subsequent availability of light for primary production. Some changes, however, may be more abrupt as the lake crosses critical ecological thresholds (e.g., step changes in the stratification regime or the loss of ice dams that retain some freshwater lakes in the High Arctic) (Wrona et al., 2006).

6.5.2.1.1. Physical effects

The most critical climate thresholds for lake ecosystems are those affecting the area and volume of standing water, while changes in the ice regime and surrounding catchments can have major impacts on aquatic habitat size and integrity and geochemical inputs (Vincent and Laybourn-Parry, 2008). In general, such changes are most apparent for relatively shallow systems, in some cases leading to their complete disappearance. For example, permafrost thawing and the production of surface to groundwater flow systems have been responsible for the elimination of many small water bodies in Siberia (Smith et al., 2005). Similarly, increased evaporative losses related to decreases in ice-cover duration (see Section 6.5.1.3) can lead to the loss of aquatic habitats, such as the drying of High Arctic ponds (Smol and Douglas, 2007b). In other regions the accelerated melting of permafrost over the past 50 years has created new basins for lakes and ponds, and increased development of shallow-water ecosystems (Payette et al., 2004; Walter et al., 2006).

The surface area and depth of lakes and ponds affect ice formation. For example, lakes shallower than about 2 m regularly freeze to the bottom (i.e., lake ice typically forms to 2-m depth in most regions of the Arctic). Two possible consequences exist if the ice regime is altered. If winters become locally warmer and precipitation (as snow) increases, then ice thickness will decrease. Consequently, habitable depths for shallow lakes and ponds on tundra previously frozen to the bottom are likely to increase and enhanced invertebrate and/or fish survival may be possible (see Section 6.5.2.1.3). Another potential consequence will be a more rapid ice loss the following spring (related to a thinner ice cover) leading to an earlier open-water season and an earlier start to spring and summer production.

For some polar lakes, ice dams from glaciers or ice shelves are the primary structures retaining the freshwater, and their collapse can result in catastrophic drainage (e.g., Mueller et al., 2003; Vincent et al., 2009). The seasonal production and melting of ice dams along the Arctic coastline are responsible for stamukhi lakes (see Section 6.5.2.3), which are important biogeochemical processing sites for large river inputs to the Arctic Ocean that may be subject to climate-related impacts in the future (Galand et al., 2008b).

Climate change is resulting in earlier dates of ice break-up (see Section 6.4.2) and, for extreme High Arctic lakes, is resulting in the onset of summer ice-free conditions in lakes that in the past have been covered by perennial ice (Mueller et al., 2009; Vincent et al., 2009). Both snow and ice affect underwater ultraviolet (UV) radiation and photosynthetically available radiation (PAR). For example, in early June in meromictic (permanently stratified) Lake A on Ellesmere Island the under-ice PAR was only 0.45% of above-ice values. Removal of snow from a 12 m^2 area on this lake resulted in a 13-fold increase in PAR under the ice and a 16-fold increase in biological exposure to UV radiation (Belzile et al., 2001). Model results suggested that such changes in snow cover would have a much greater effect on underwater UV exposure than moderate stratospheric ozone depletion (Vincent et al., 2007). In some ice-covered lakes, much of the photosynthetic production in the water column is associated with a deep maximum of phytoplankton or photosynthetic sulfur bacteria. In Lake A, past changes in planktonic production as inferred from pigment concentrations in sediments have been attributed to climate-related changes in snow and ice cover (Antoniades et al., 2009).

For some lakes, the loss of ice can result in the loss of vertical habitat structure and cooling (Vincent et al., 2008a). As part of the modeling of broad spatial patterns of changes in lake ice noted in Section 6.4.3, Dibike et al. (2010) evaluated changes in water temperature and thermal structure in lakes across the Northern Hemisphere. For example, Figure 6.18 shows the mean annual cycle of simulated water-temperature profiles in hypothetical lakes of 20-m depth along longitudinal transects at 105° W and 90° E, representing cross-sections through central continental areas of North America and Asia, respectively. For the two selected profiles, higher-latitude lakes along 105° W show less summer stratification than those along 90° E, which could be due to differences in relative coldness and/or elevation of the two regions. As noted by the authors, the projections are only intended to evaluate broad spatial patterns, and more site-specific modeling of lake types at higher spatial resolution is required to probe such regional differences. In general, however, results suggest that future warming will result in an overall increase in water temperature, with summer stratification starting earlier and extending later into the year.

Warming of the underlying water column by radiation is controlled to varying degrees by the thickness and composition of the snow- and lake-ice cover; white and black ice, for example, have different levels of albedo and transmissivity. In combination with water color and transparency, this affects heating rates, depths and mixing (e.g., Cahill et al., 2005). Earlier thinning and loss of ice cover also contribute to enhanced heating of the water column, which ultimately sets conditions for earlier and shallower development of the thermocline by increasing temperature differentials between surface and bottom waters. Longer open-water periods can further enhance overall lake warming, the combined effect being to drive high-latitude lakes from monomixis (a single period of mixing each year) to dimixis (the water column stratified for part of the summer). This threshold effect as a result of water temperature rising above the point of maximum density (~4 °C), can affect many other habitat properties, such as nutrient regimes and water-column oxygenation (see also Section 6.5.2.1.3).

6.5.2.1.2. Chemical effects

The unproductive lakes of the Arctic strongly reflect processes operating in the contributing catchments. For example, direct linkages between terrestrial processes and dissolved organic carbon (DOC) concentrations in lake waters are well known. Terrestrial DOC plays a key role in aquatic ecosystems as it affects primary and secondary productivity, community structure and metabolic balances, availability of dissolved nutrients and metals, and the thermal structure and optical properties of water bodies. Temporal

changes in DOC concentrations have been attributed to changes in runoff, temperature, solar radiation, soil moisture, growing season length, and atmospheric deposition chemistry (Williamson et al., 1999 and references therein). Changes in the seasonal variability can also have far-reaching biological consequences (Weyhenmeyer, 2009a). In Sweden, for example, water chemistry changes in lakes are still more pronounced south of the Arctic (Weyhenmeyer, 2008, 2009b), but this could change in response to the accelerated thawing of permafrost (see also Chapter 5).

Lake ice has a controlling influence on oxygen conditions, and even moderately productive lakes can be driven to anoxia during winter ice conditions (e.g., Laurion et al., 2010). This in turn affects a great variety of biological and biogeochemical processes. The shift of lakes from cold monomictic (continuous mixing in summer) or polymictic (multiple episodes of mixing in summer, favored by cold temperatures; e.g., Lac à l'Eau Claire in Nunavik, Canada; Milot-Roy and Vincent, 1994) to dimictic (stratified in summer) will increase the possibility of oxygen depletion and even anoxia in the bottom waters during their periods of summer stratification, thereby reducing habitat availability for high oxygen-demanding biota such as Arctic char (*Salvelinus alpinus*). Conversely, the corollary of such conditions is a shorter duration of winter ice cover, which acts as a barrier to oxygenation from the atmosphere and prevents wind-induced mixing. In northern temperate lakes, severe oxygen depletion under the ice can lead to the 'winter kill' of resident fish. This is likely to be reduced in a warmer climate with reduced ice duration, with potential cascading effects on lower trophic levels (e.g., Balayla et al., 2010).

Thermokarst or thaw lakes, a major class of high-latitude aquatic ecosystems, are also sensitive to shifts in oxygen. Lake ice affects the concentration of greenhouse gases in two ways: by acting as a physical barrier to efflux to the atmosphere, and by modifying oxygen tension in the water column and sediments that in turn affect the balance of methanogenesis (production of methane) versus methanotrophy (breakdown of methane). The production of methane (methanogenesis) by microbes in lake sediments is an anaerobic process favored by anoxia. Anoxia is common year-round in organic-rich lake sediments and within centimetres of the sediment-water interface due to microbial oxygen consumption that exceeds the rates of oxygen diffusion through the water column and sediments. Anoxia in lake bottoms is enhanced during summer and winter stratification. These effects are receiving increasing attention in permafrost thaw lakes (thermokarst lakes and ponds) (Vincent and Laybourn-Parry, 2008). Water-column profiles and surface sediment analyses of dissolved gases (carbon dioxide and methane) in 34 shallow monomictic and deeper dimictic thermokarst lakes of the Kolyma lowlands during 1994, 2002, and 2010 showed anoxic conditions and high methane concentrations prevalent in surface sediments of all lakes (Zimov et al., 2001). Concentrations of dissolved gases were significantly elevated in the bottom waters of stratified Siberian thermokarst lakes, a phenomenon that has long been observed in non-thermokarst lake systems (Rudd and Hamilton, 1978; Michmerhuizen et al., 1996). Breton et al. (2009) also observed widespread supersaturation of methane in 46 Nunavik lakes regardless of mixing regimes. Studies in the Nunavik thaw ponds show that some are highly stratified and likely to be producing methane in their anoxic bottom waters for almost all of the year (Laurion et al., 2010).

Methane production in thermokarst lake sediments continues throughout the winter ice-cover period (Walter et al., 2006, 2008a). Spring ice melt in thermokarst and non-thermokarst lakes releases large quantities of methane and carbon dioxide to the atmosphere via diffusion (Michmerhuizen et al., 1996; Phelps et al., 1998) and ebullition (Walter et al., 2008a). In winter, methane bubbles released from the lake bottom by ebullition become entrained in lake ice as it thickens at the surface (Walter et al., 2006; Figure 6.19). Some gas escapes during winter as bubbles move beneath the ice to cracks or open-hole hotspot seeps, where vigorous rates of bubbling can maintain relatively ice-free holes year round. However, large volumes of methane are trapped and stored in lake ice throughout winter and released to the atmosphere during spring ice melt (Walter et al., 2008a). The bubbles in the ice provide information on the location of discrete methane-seep production in lakes and can be detected by SAR, which may provide a remote sensing approach to scaling up estimates of methane fluxes to the regional and circumpolar scale (Walter et al., 2008b).

Loss of ice cover and associated enhanced warming of these permafrost lakes could greatly increase methane production from the vast wetland-lake regimes of northern latitudes (Section 6.1.3) (Nozhevnikova et al., 1997; Metje and Frenzel, 2007). However, thermodynamic controls over biological methane production also affect methane oxidation, commonly an aerobic (oxygen-requiring) bacterial process that converts methane to the less potent greenhouse gas carbon dioxide (Rudd et al., 1974) and enhances oxygen depletion under ice (Rudd and Hamilton, 1978). The net balance of the two processes in scenarios of future warming in lakes is uncertain, although short-term sediment warming experiments on Swedish lake sediments suggested that temperature sensitivity was a stronger control over methane production than oxidation, which was governed by substrate availability (Duc et al., 2010). This implied that elevated temperatures will enhance methanogenesis, which may cause increased methane release from sediments until methane oxidation increases in response to higher methane levels. The end result is difficult to predict, especially as changes in the oxygen dynamics of high-latitude lakes induced by ice-related shifts in water budgets and levels, stratification, and mixing could also affect the relative balance of methanogensis and methanotrophy. Relevant greenhouse gas studies of these types of lentic systems include those by Walter et al. (2006, 2007a,b, 2008a,b), Breton et al. (2009), Duc et al. (2010), and Laurion et al. (2010).

Ice cover and the associated thermal stratification (Section 6.5.2.1.1) play a significant role in the fate of contaminants supplied to northern lakes, such as by stream inflow or direct atmospheric deposition. In the case of inflow, thermal structure is known to affect the amount of contaminant retention. For example, in Amituk Lake of the Canadian High Arctic 59% of the annual mercury load flowing into the lake traversed the lake as a mercury-rich buoyant current immediately beneath the ice, ultimately leaving the lake via its outflow. Only later in the summer, when the ice had disappeared and the water column had warmed to become isothermal, did the inflowing mercury and other solutes mix into the lake water (Semkin et al., 2005). The inflow of contaminants under changing ice conditions is of particular concern because their inflow to lakes might be enhanced by other cryospheric changes on the landscape (e.g., thawing permafrost and melting of snow and glaciers) that result in the release of historically deposited contaminants (e.g., Klaminder et al., 2008; Faïn et al., 2009). In the absence of lake inflow, ice cover and associated thermal stratification may still play a role in influencing the vertical distribution of contaminants.

Changes in ice cover are also likely to affect the fate of volatile chemicals deposited directly from the atmosphere. Lake sediment records are commonly used to investigate trends in contaminant levels (Stern et al., 2005; Outridge et al., 2007; Muir et al., 2009) and carry the potential to distinguish between possible climatic effects and changes in atmospheric deposition of (mostly volatile) contaminants to lakes and surrounding catchments. In general, an expansion of the ice-free period will result in a shorter period during which contaminants can re-volatilize from the snow and ice cover but a longer period for direct deposition to the lake water. For photosensitive contaminants that enter ice-covered lakes, alterations to the light regime produced by changes in ice-cover thickness and composition (see Section 6.4.3) could alter their chemistry (e.g., Hammerschmidt et al., 2006), although little is known about the role played by photosensitivity in the overall cycling of contaminants. In general, in-lake production, metabolism, and (photo-)degradation rates of certain contaminants may be enhanced by longer ice-free periods (Hammerschmidt et al., 2006).

Opening of ice-free zones and increases in water temperature from the loss of ice cover may also affect in-lake processing of contaminants. Greater methylation of mercury, for example, is likely to result from higher temperatures, particularly in shallow zones. Higher water temperature is also likely to increase pelagic production and thereby enhance algal scavenging of mercury, a proposed entry pathway for mercury to food webs (Outridge et al., 2007). Overall, higher water temperatures associated with a decrease in ice cover, and related changes in food and energy pathways and/or productivity (benthic to pelagic), are likely to modify contaminant transfer through such lakes (Chételat and Amyot, 2009; Carrie et al., 2010; Gantner et al., 2010a,b).

6.5.2.1.3. Biological effects

Ice is a key physical parameter that both structures and regulates abiotic and biotic processes within Arctic aquatic ecosystems. Biotic responses are induced at an individual, population, or community level depending upon the nature (rate, direction, magnitude, spatial scale) of abiotic change. Accordingly, shifts in ice characteristics will cascade through ecosystems, resulting in widespread alterations. For example, in addition to effects at the individual level (e.g., displacement from preferred habitat, alteration in growth rates) and population level (e.g., changes in distribution and abundance), changes in ice will affect trophic coupling, potentially engender mismatches between physical drivers and biotic responses, and affect phenological events such as the timing of key system transitions and life history shifts in biota.

In the case of photosynthetic production in lakes, the duration of open water is particularly critical. For example, the 250 000-year paleolimnological record from Lake El'gygytgyn, an ancient crater lake in the Siberian Arctic, showed that periods of the highest primary productivity were associated with warm, ice-free summer conditions, while the lowest rates were associated with periods of perennial ice coated by snow (Melles et al., 2007). In addition to improved light conditions for photosynthesis, measured levels of primary productivity could be further compounded by alterations in other environmental factors, such as increased wind-induced mixing and entrainment of nutrients into the euphotic zone (the surface layer with enough light for net photosynthesis), and catchment geochemical inputs.

Changes in the timing of freeze-up and break-up on lakes will also affect important biological aspects of seasonality, which is defined as a predictable change or pattern in a time series that recurs or repeats over a one-year period. For example, the seasonal succession of plankton has been well described in many north temperate lakes and is strongly coupled with the freeze-up and break-up of ice cover and summer thermal stratification (Sommer, 1989). A variety of structural and functional ecosystem changes in such lakes have been coupled to the changes in seasonality, in particular to an earlier ice break-up and an earlier onset of stratification, and provide insights into how Arctic lakes may respond. One of the most obvious effects of an earlier timing of temperate lake-ice break-up has been an advanced spring phytoplankton bloom (Weyhenmeyer et al., 1999; Gerten and Adrian, 2002; Winder and Schindler, 2004; Peeters et al., 2007) often resulting in an earlier zooplankton biomass peak (Straile, 2000; Gerten and Adrian, 2000). However, a synchronous response to these higher spring temperatures is usually restricted to fast-growing plankton, while slow-growing species with complex life histories show species-specific responses (Winder and Schindler, 2004; Adrian et al., 2006).

Changes in lake-ice regimes will have significant impacts on primary productivity and related trophic relationships in Arctic lakes. For example, increased temperatures and stratification associated with decreases in ice cover, accompanied by larger nutrient inputs, may favor the development of certain phytoplankton. In the case of noxious blooms of cyanobacteria, this could be a significant concern. Seasonality of the plankton is also likely to be affected by temporal changes in ice coverage given that flagellate plankton have been observed to be abundant below the ice in Arctic lakes, whereas diatoms appear once ice is gone. In general, although photosynthesis does take place beneath an ice cover, it is expected that primary production will increase with decreased ice thickness and snow cover (e.g., Vincent et al., 2008c). Evidence for this is provided by present conditions where snow-free ice conditions can lead to bloom concentrations of photosynthetic flagellates (Weyhenmeyer et al., 1999). However, in Arctic regions projected to experience increases in surface accumulations of snow and/or the formation of white ice (see Section 6.4.3), under-ice plankton abundance could be negatively affected. Such changes in snow and white-ice coverage are also likely to affect levels of secondary productivity. Fish production in northern alpine lakes, for example, has been linked to snow depth (e.g., Borgstrøm and Museth, 2005; Prowse et al., 2007b).

Changes in water-column stratification associated with increased duration of open water can potentially result in the loss of some species and the establishment of others. For example, the diverse, highly stratified communities of single-celled Archaea in High Arctic lakes are likely to be disrupted by future

changes in ice cover (Pouliot et al., 2009). By contrast, increased open water can allow the development of new trophic levels and even the establishment of aquatic bird species (Vincent et al., 2009). Warmer, more nutrient-rich dimictic conditions may also favor cladocerans (see Sorvari et al., 2002). Importantly, this could result in increased bioaccumulation of methylmercury relative to copepod-dominated zooplankton communities (Chételat and Amyot, 2009), with the potential for increased mercury transfer to fish and humans, although effects of biodilution may counteract this effect (Gantner et al., 2010a).

Depending upon latitude, as well as lake characteristics such as depth, morphometry, and elevation, early thermocline development (Section 6.5.2.1.1) will profoundly alter lake ecosystems as well as the cold-water fish species present. Using lake trout in North America with temperature preferences of 10 to 12 °C as an example, under climate warming southern (southern boreal) lakes will experience earlier and perhaps deeper thermocline formation than at present. Accordingly, the metalimnion (middle layer of a thermally stratified lake) and hypolimnion (lower layer of a thermally stratified lake) volumes will be smaller. These areas are used as a summer thermal refuge by lake trout (*Salvelinus namaycush*) at southern latitudes to escape epilimnion (upper mixed layer of a lake) temperatures of more than 12 °C; thus, smaller volumes of preferred habitat will lead to stress for individuals, likely to be associated with population consequences such as lowered growth and/or a decline in numbers. In mid-latitude lakes (e.g., 60° N, northern boreal), lake trout tend to use the upper layers of water, which more closely approximate their thermal preferences and thereby promote growth. Projected effects of earlier ice loss and radiative heating suggest that shallower and earlier thermocline development (which will decrease the availability of this habitat) will have parallel consequences to those described above for more southern lakes (Mackenzie-Grieve and Post, 2006), despite the habitat being used differently. At least over the near future, suitable thermal habitats for lake trout in Arctic lakes are likely to remain similar to those at present or increase in volume, thus promoting lake trout growth (provided that all other factors are equal). The key environmental drivers producing this scenario are shifts in the timing of lake-ice melt combined with a longer heating season.

The effects of wind are a complicating factor in ice dynamics and thermal structure affecting habitats. Earlier ice loss results in larger fetches being open earlier and longer. Wind-driven mixing of surface waters will almost certainly interact with heating to complicate thermocline development and depth; however, the nature of such effects remains unclear. Although general logic models and trends connecting ice loss to fish populations can be developed, high local and interannual variability of effects will be very likely. Given that lake trout are generally long-lived, such climate change signals may not be readily discernible within the populations. Moreover, decreased occurrences of winter fishkills due to oxygen depletion events will generally be an additional effect of reduced ice-cover duration (Stefan and Fang, 1997). However, the significance of this is likely to vary by latitude and lake characteristics.

Ice cover also affects the migration and dispersal of aquatic organisms. A small number of Arctic lakes are permanently ice-covered (e.g., Vincent et al., 2008a), and in hydrological terms these are poorly connected to their surroundings. In these lakes, melt-out is typically restricted to a narrow moat. This greatly limits the wind-induced mixing of inflowing stream water and limits the presence of some biota, such as aquatic birds. Increased melting of ice and snow in both the catchment and lakes in a warmer climate may result in an increased overflow and, consequently, a greater hydrological connectivity between the lakes (Kusumastuti et al., 2008; see also Figure 6.1c). Conversely, changes in ice jams in large river ecosystems, particularly in their delta environments, could cause less frequent flooding of higher elevation lakes that rely on such events for water, sediment and nutrient supply (e.g., Prowse et al., 2002b; Peters and Prowse, 2006; Lesack and Marsh, 2007) (see Section 6.5.1.1 regarding physical sensitivity of delta systems to changes in flooding regimes and Section 6.5.2.3 for biological effects).

6.5.2.2. Lotic systems

6.5.2.2.1. Physical effects

In Arctic rivers, ice is important in defining the in-stream habitat for fish, invertebrates, and aquatic plants (Prowse, 2001a; Huusko et al., 2007), and climate change will have a profound impact on the future ice regime (Section 6.5.1.1). Various forms of river ice will directly create important habitats and will alter river geomorphology through erosion and sedimentation processes (Section 6.5.1.4). River ice will also influence the behavior and biological response of stream biota (Huusko et al., 2007) and, therefore, will play a central role in their growth, survival and reproduction.

Surface ice creates shelter habitats for fish in areas that are too exposed for use during open-water periods (Stickler et al., 2007; Linnansaari et al., 2009). A reduction in such ice shelter will, therefore, lead to a loss of suitable winter habitat during the period when the water temperature has not yet reached the level to cause changes in fish habitat use. In small and steep streams, winter formation of ice will define habitat availability and distribution independently of changes in discharge (Stickler et al., 2010). In such environments, a shorter ice season will influence habitat diversity. In addition, with future climate warming, an increased number of winter warm spells leading to mid-winter ice break-up may have a significant influence on habitat availability. Many Arctic rivers that currently have bed-fast ice, and thereby no available winter habitat, may shift into a regime with a floating ice cover. This will create new habitat for winter survival of species in these rivers.

By altering geomorphology and vegetation in Arctic rivers, river ice processes, particularly ice break-up, have important effects on channel development and maintenance (Section 6.5.1.4). A reduction or disappearance of these processes in the future may lead to degradation of existing habitat or changes in habitat composition over time. This will influence biological productivity and biodiversity over the long term (Prowse, 2001b).

6.5.2.2.2. Chemical effects

Long periods with complete ice cover can decrease dissolved oxygen (DO) concentration in river water to a level harmful for the survival of invertebrates and fish (Prowse, 2001a). Several factors influence the decline in DO after the complete freeze-over of a stream. Of particular importance are the water-column and riverbed oxygen demands (Prowse, 2005). In rivers dominated by autumn flows of low-level DO, the decline from loss of air-water re-aeration can be particularly important. Further development of DO over winter is dependent on the source and chemical properties of winter discharge, residence time, and influx of oxygen-consuming material. It is important to note that changes in DO of ice-covered rivers do not follow a general pattern but are a complex product of a number of factors (Prowse, 2005). Changes in winter-season length, ice thickness, and the number of zones with broken ice cover could allow species sensitive to low DO concentrations to utilize larger habitats in the river and thereby increase survival (Wrona et al., 2005). Changes in ice cover may also strongly influence nutrient cycling in Arctic rivers. Greenwald et al. (2008) and Zarnetske et al. (2008) concluded from fieldwork conducted on streams on the North Slope of Alaska that climate warming will not significantly expand the thickness of the hyporheic zone – a zone of benthic substrate through which stream water readily flows, found in most Arctic rivers. The thaw basin under Arctic streams flowing over permafrost may increase, but the hyporheic zone currently only occupies the upper part of the thaw basin and might not deepen substantially in the future. However, the season of flowing water in the Arctic is expected to lengthen substantially, and thus the opportunity for nutrient and organic matter processing in the hyporheic zone will also be modified. This shift in the seasonality of nutrient processing may have important impacts on biota in Arctic streams and rivers.

Currently the spring break-up of river ice causes significant bank erosion that delivers large quantities of sediment and nutrients to streams and rivers (Prowse et al., 2006). Increased nutrient loading might be considered a benefit if it stimulates additional productivity in streams but could also be a problem if it alters the structure of the benthic biotic community (e.g., Bowden et al., 1999; Prowse, 2001b). Furthermore, increased sediment loading is likely to impair stream biota by increasing turbidity, which will decrease light penetration (and thus decrease primary production), increase abrasion of sensitive

biota, and when accumulated in benthic habitats, restrict access to interstitial spaces in substrate and reduce intra-gravel flow. If the spring break-up intensity decreases in the future, then less sediment and fewer nutrients may be delivered to rivers during the relatively short spring freshet.

However, greater variability in future temperature and the form (rain or snow) and intensity of precipitation will lead to increased surface and bank erosion (e.g., Lamoureux and Lafreniere, 2009; see also Chapter 5). This may increase sediment transport to streams and the load of associated nutrients. Nutrient and sediment loading will be especially large if future warming increases the formation of thermokarst features that directly affect water bodies. Bowden et al. (2008) and Gooseff et al. (2009) concluded that, in the long term, the negative impacts of increased sediment load would outweigh the positive effects of increased nutrient loading. Thus, future warming in the Arctic is likely to alter water quality in ways that could be detrimental to some macroinvertebrates and fish living in Arctic rivers and streams.

6.5.2.2.3. Biological effects

Changes in the seasonality of ice in Arctic streams will alter the timing and magnitude of sediment and nutrient delivery. Such changes will be driven largely by increasing temperature and will affect terrestrial and aquatic environments. Most lotic environments in the Arctic depend on or are limited by allochthonous (material originating from outside the water body) inputs from the landscape, making terrestrial-aquatic linkages important factors. In addition, terrestrial environments may respond differently to changes in light than aquatic environments. The life-cycle of higher vascular plants is driven by daily and seasonal photoperiods. Since the photoperiod will not change, short-term responses by such terrestrial plants to a longer potential growing season will be limited; however, long-term changes may include shifts in species composition. By contrast, productivity and growth of microbial communities in soils and algae in streams and rivers are related to the photoperiod to a lesser extent. Such communities are often limited by light and/or temperature, and respond readily to changes in these variables as well as to changes in nutrient supply. Thus, aquatic microbial communities and algal assemblages may track changes in light and temperature (as a result of reduced ice cover). In-stream productivity or nutrient demand may thus increase, creating or intensifying asynchrony among nutrient delivery/allochthonous inputs (regeneration in soils by microbes, nutrient uptake by terrestrial plants, and surface and subsurface nutrient leaching to streams) and nutrient use and recycling within streams. In turn, this asynchrony in seasonal patterns of nutrient production and use may cause a shift in species composition and productivity in streams and rivers. Eventually, the terrestrial environment may adapt to the warmer climate through succession, but at this time it is not clear whether or how a modified terrestrial ecosystem in the Arctic will affect aquatic ecosystems.

Changes in ice cover can also have a direct impact on fish productivity and mortality. For example, for Atlantic salmon (*Salmo salar*) adapted to complete ice cover, removal of an ice cover has been shown to produce significant negative effects on their energy budget (Finstad et al., 2004a). Energy deficiency is important to winter survival, and a change in ice cover can reduce their ability to survive winter (Finstad et al., 2004b). Anchor-ice formation on the bed of streams also influences habitat availability and use. The response of fish to anchor ice is dependent on the severity of the ice formation, with related behavior spanning from avoidance to active use (Huusko et al., 2007). Periods with heavy formation of anchor ice can lead to increased movement in fish (Stickler et al., 2007), which is considered a problem due to increased use of energy during the winter period. The future development of anchor ice will depend on local climate and flow conditions. Longer freeze-up periods with higher flow, for example, could extend the period during which anchor ice develops. Winter break-ups and subsequent re-freezing could also increase the number of anchor-ice formation periods. Increased anchor-ice formation and release can also enhance anchor ice-induced sediment transport (Prowse, 2001a), which could have additional negative biological consequences for fish and benthic biota.

Movement of salmonids to overwintering habitats mostly occurs prior to ice formation in rivers; however, local movements between habitats also occur after ice formation (Jakober et al., 1998; Linnansaari et al., 2009). Extensive anchor ice precludes access, whereas patchy anchor ice and ice-covered areas appear to be preferred (Linnansaari et al., 2009). Accordingly, reduced ice cover or duration on river systems is likely to result in a tradeoff between increased habitat (or access to such) with that habitat being less preferred due to lack of surface ice cover. Access, primarily by migratory anadromous salmonids, to key overwintering habitats may thus provide benefits to overall population survival and productivity.

The Arctic Climate Impact Assessment identified changes in break-up timing as one of the greatest freshwater-ice effects of a changing climate (Wrona et al., 2005). A concern particularly identified was the development of mid-winter ice break-ups (Section 6.5.1.1), an event poorly studied in regard to impacts on aquatic ecosystems, particularly those at high latitude. Based on research in a more temperate, southern environment, however, Cunjak et al. (1998) demonstrated that mid-winter ice break-ups may have large impacts on the survival of different life stages of Atlantic salmon and that more frequent mechanical winter break-ups could result in increased mortality. The ice-scour damage associated with such events will also influence the composition of riparian and aquatic vegetation, typically leaving elevational tiers of vegetation type corresponding to break-up frequency and severity (Prowse and Culp, 2008). Hence, in a situation with reduced break-ups, the ice-controlled vegetation pattern is likely to diminish or disappear, producing a more monoculture environment. By contrast, more mechanical break-ups will increase the level of vegetation disturbance and removal.

Large-scale changes in ice-cover duration and break-up timing will alter flow regimes and thereby influence Arctic rivers as migratory routes, affecting the timing of fish runs or even the migration of large mammals such as caribou (Sharma et al., 2009). Changes in flow timing in spring will also influence conditions for fish out-migration (Reist et al., 2006a,b). The loss of an ice cover is likely to increase the risk of predation on stream-living animals from mammalian and avian predators due to the loss of critical in-stream shelter. Moreover, winter and related ice formation could act as a 'bottleneck' for survival of fish and invertebrates. Results reported by Huusko et al. (2007) suggest that the variability in creating such bottlenecks among rivers is highly context dependent and controlled by the life stage of the fish, local habitat, and the related type of ice regime. Overall, the large and complex scope of potential changes in future river-ice regimes outlined in Sections 6.5.1.1 and 6.5.1.2 will make predictions of future biological responses difficult, particularly considering the current rather limited knowledge of high-latitude lotic systems.

6.5.2.3. River delta ecosystems

A number of major river deltas are found along the Arctic coast as well as on the rivers that drain into the Arctic Ocean. Major examples of the former include the Yukon, Colville, Indigirka, Kolyma, Lena, and Mackenzie river deltas, and of the latter the Peace-Athabasca and Slave river deltas (e.g., Prowse et al., 2006); both the Lena and Mackenzie are considered to be mega deltas (Anisimov et al., 2007). Much of this area, however, is a riparian environment composed of numerous distributaries and a multitude of small basins connected by varying degrees to the channel network.

The annual flow and water-balance regimes of these delta systems are dominated by the spring freshet, and although driven by discharge arising from southern headwaters, it is the effects of river-ice break-up that produce the highest flood stages (e.g., Goulding et al., 2009a,b). It has been recognized for some time that the water budget and nutrient-sediment supply of delta riparian zones are heavily dependent on ice-jam floodwaters (see Section 6.5.2.1.3 and Figure 6.1c). The strength of this dependence has been reinforced by recent work on the Mackenzie River Delta, which contains about 45 000 riparian lakes. Specifically, decreases in the severity of river-ice break-up has lessened the flooding of high closure lakes, which has the potential to result in the loss of some of these water bodies and changes in the biogeochemical processing of river water by the floodplain ecosystem (Figure 6.20) (Lesack and Marsh, 2007). Hence, as discussed in Section 6.5.1.1 and as found for another delta in the headwaters of the

Mackenzie River (the Peace-Athabasca Delta), future climate conditions that produce thinner ice and reduced spring runoff (due to a smaller winter snowpack) will lead to overall reductions in ice-jam flooding (Beltaos et al., 2006). This could pose a major threat to the health of riparian ecosystems, both for resident biota and for those that use them for migratory purposes, such as waterfowl.

Reductions in the erosional forces of river ice as a result of decreases in break-up intensity (Section 6.5.1.4) are also likely to alter the extent of particle release from riparian zones to lakes and rivers and affect riparian vegetation communities (e.g., Prowse, 2001b). Given the importance of particles for microbial colonization and biogeochemical processes in many northern waters (e.g., Galand et al., 2008a), as well as for underwater light attenuation (e.g., Retamal et al., 2008), this could result in changes in biogeochemical processing rates, primary production, and greenhouse gas fluxes (Vallières et al., 2008).

Freshwater environments at the coast are also prone to the effects of changing ice conditions. In the High Arctic, for example, fjords can be blocked by thick multi-year sea ice and ancient ice shelves, resulting in an extensive layer of freshwater called an 'epishelf lake'. These environments are proving to be microbial ecosystems with diverse biological communities; however, they are extremely vulnerable to ongoing climate warming and the loss of ice. Observations along the northern coastline of Ellesmere Island, Canada, have shown that many of these unique ecosystems have been driven to extinction as a result of recent climate change, and that they are sensitive indicators of climate change (Veillette et al., 2008). Other ice-dependent freshwater lakes at the High Arctic coast are becoming inundated with seawater as a result of the loss of integrity of their retaining ice dams (e.g., Vincent et al., 2009), and the extensive, microbiologically rich ice-bound lakes on ice shelves are disappearing completely as a result of their melting and collapse (Mueller et al., 2008).

Stamukhi lakes are another important class of coastal freshwater system that have implications for the marine environment. These ephemeral lagoons occur throughout winter, spring, and early summer along the Arctic coastline and form behind pressure ridges ('stamukhi') of thick, partially grounded sea ice near large river inflows. Recent studies on the stamukhi lake, Lake Mackenzie (Figure 6.21), have revealed it to have an active microbial ecosystem with distinct physical and microbiological properties. This type of circumpolar ecosystem is likely to play a key functional role in processing riverine inputs to the Arctic Ocean (Galand et al., 2008b), which could be significantly altered by the effects of climate change (Dumas et al., 2006).

6.5.3. Socio-economic consequences and adaptation options

- Transportation and hydroelectric production are two of the socio-economic sectors most vulnerable to change in freshwater-ice regimes; ice roads are currently a vital link for industry and northern communities.

- Continued warming will preclude ice roads as a major form of northern transportation; alternative forms of transportation will be needed, but the capital costs of these are likely to be enormous.

- Changes to ice regimes will make the practice of some traditional subsistence-based lifestyles potentially hazardous and may reduce the ability to undertake some traditional harvesting methods.

- Hydroelectric operations will both benefit and be challenged by changes in river-ice conditions; the potential for winter break-ups is likely to pose operational challenges to ensuring that flow releases do not cause damage downstream.

- Monitoring and mitigation of ice-related problems will be a particular issue for hydropower producers with remote facilities.

- The importance of river ice to hydroelectric operations may be affected by future energy adaptations involving renewable but non-storable energy sources that require less regular hydroelectric operations to maintain a continuous energy supply.

6.5.3.1. Northern infrastructure, transportation, and traditional lifestyles

Lake and river ice provide seasonal transportation platforms throughout the Arctic. Many northerners depend on this natural network for access to hunting, fishing, and reindeer herding or trapping areas, often in support of traditional subsistence-based lifestyles (e.g., Furgal and Prowse, 2008; Prowse and Brown, 2010). Changes in ice regimes, however, will make such access more uncertain and potentially hazardous, for example, due to shorter ice duration, thinner and less stable ice regimes, and mid-winter thaws. Moreover, these changes may reduce the ability to undertake some traditional harvesting methods. For example, lake and river ice have been used as a stable and long-term platform for the deployment of specific fishing gear (e.g., gillnets); the loss of these platforms will require adaptation of traditional practices (Reist et al., 2006c). Ice-based travel also provides the principal transportation route for some isolated communities and industrial developments (Vuglinsky and Gronskaya, 2006; Prowse et al., 2009a).

Open-water transport on rivers and lakes has been the historical method of transport to many northern population and industrial centers, particularly along major northward-flowing rivers. These transportation links are so important that nuclear-powered ice breakers are used on Russian rivers to expand the shipping season, such as along several hundred kilometres of the Yenisey River between Igaraka or Dudinka and its northern outlet. Any increase in the ice-free season from climate warming will reduce the costs of such operations and enhance south-north and north-south transport, the latter becoming increasingly important as the Arctic becomes an expanding source of resources. For example, a six- to nine-week reduction in the ice season on the Mackenzie River could result in a 50% increase in the use of barge-based transport (Lonergan et al., 1993). However, in the case of major lake and river ice-based transportation systems, such as the complex terrestrial and water network built around the inland icebreaker system of Russia, any projected changes in the ice season would require major coordination challenges in formulating new schedules or constructing new storage terminals and relay points.

Beyond the main rivers, there also exists a winter ice-road network composed of a combination of private and public lake and river crossings that link all-season road systems, communities, and remote industrial and mining complexes. Such networks are especially important for parts of northern Canada, such as the Northwest Territories, where the public roads system almost doubles during winter, or the vast territory of Nunavut, where there are no long-distance all-season highways (Prowse et al., 2009a). Similarly, annual ice roads are also constructed across vast areas of the Russian North primarily north of 60° N but extending from the Kola Peninsula to as far as the Bering Sea coast.

Although scientific publications that explicitly detail the importance of ice roads to northern communities are rare (e.g., Ford et al., 2008), there are many accounts in the public press when such networks are affected by unseasonably warm weather. To illustrate the significance of these events, one recent example is described here based on accounts provided by Carlson (2010). Specifically, mild weather in March 2010 caused the province of Manitoba, Canada, to close a 2200-km winter road network composed of muskeg (bogland), lakes, and rivers. The road system had deteriorated to the point of stranding numerous freight haulers and local drivers on thawed winter roads, necessitating emergency evacuations. Typically, the road carries more than 2500 shipments each year to more than 30 000 First Nations people. In response to dwindling construction supplies, rising food and fuel prices, and a related rise in unemployment, First Nations Chiefs declared a state of emergency in eleven communities. Carlson et al. (2010) also noted that because of deteriorating conditions approximately 600 km of the winter road system have been relocated to land since 2001 (Government of Manitoba, 2010) and spending on winter roads has tripled since 1999 (Government of Manitoba, 2009).

Ice roads are also critical to the resupply of the complex of mining centers, which cannot use air access for the transport of heavy loads, fuel, and large equipment. One example is the 600-km long Tibbitt to Contwoyto Winter Road in northern Canada, which travels over 495 km of frozen tundra, lakes, and rivers (Figure 6.22). Although it typically operates for only two months per year (February and March), at a approximate annual cost of CAD 10 million, it carries up to 8000 truck loads per year, each weighing an average of 30 tonnes, with the load capacity rising as the ice thickens and increases in bearing strength. It has been estimated to contribute significantly to the territorial and national annual economies – approximately CAD 800 million and CAD 350 million, respectively, in 2001 but rising significantly with enhanced northern development (EBA Engineering Consultants, 2001). A similar example is the 360-km long winter road in the Chukotka region of Russia constructed each winter from the ocean port Pevek, over tundra, lakes, and streams to the Kupol gold and silver mine at Bilibino (Noble, 2009). In such cases, reductions in ice duration, thickness, or mechanical strength (e.g., related to changes in the amount and/or timing of snow-cover loading) could have major implications for such remote developments. For some Arctic centers, changes in ice-related transport can have both positive and negative effects. In the case of Arkhangelsk on the Northern Dvina River, Russia, an increased shipping period and freight turnover from the inland navigation fleet would result from a decrease in ice duration, but, on the other hand, delays in the building of ice-road crossings would create substantial difficulties for local freight and public transportation (e.g., Ginzburg, 1989).

Initially, adaptation to the reduction in the size of maximum loads that can be safely transported on northern ice roads could involve (i) modifications to techniques involved in ice-road construction, such as by enhanced surface flooding or spray-ice layering; or (ii) where transport capacity is not already maximized, modification of transport schedules to concentrate more on the coldest part of winter (Prowse et al., 2009a). Continued warming will preclude ice roads as a major form of northern transportation, and there will be a need for alternative forms of transportation. In cases where an open-water network is feasible, transport by barge could be possible. For land-locked locations, however, the only viable option for heavy-load transport will be the construction of land-based road or rail networks. The initial capital costs of these, however, are likely to be enormous, especially where they must pass over terrain that is also projected to experience significant permafrost thaw and subsidence from climate change.

Although relatively rare, there is also some infrastructure within the Arctic that is located in river channels, such as the developing gas fields of the Mackenzie River Delta. The flood-damage exposure of such facilities depends on how climate will affect the severity of ice-jam and related backwater flooding. For coastal locations, the effect of sea level rise must also be factored into assessing such flood risks (see Section 6.5.1.1). One method identified to avoid these potential impacts involves the use of river barges for production facilities as opposed to being entirely land based (Prowse et al., 2009a).

6.5.3.2. Hydroelectric power

Production of hydroelectric power is important in several Arctic countries, the operations of which are seasonally constrained by the effects of river ice that could markedly change under future climatic conditions. At present, the total installed capacity in the Arctic countries (2006 data) is approximately 80 GW (Figure 6.23), but for many areas, unregulated large northern rivers still hold vast potential (e.g., Prowse et al., 2004, 2009a). With the future projections of inflow, this potential will probably increase for most of the Arctic region (Hamududu and Killingtveit, 2010).

Changes in ice conditions can affect hydroelectric operations in a number of ways, both positively and negatively. For example, the estimation of ice loads on facilities such as dams, intakes, outlets, and gates is important both for engineering design and operations (Comfort et al., 2003). A shorter ice season and thinner ice cover (Sections 6.4.3 and 6.5.1.1) could reduce the static ice loads on dams, but on the other hand, a more unstable winter with mechanical ice break-ups could increase the dynamic loads on in-channel facilities. More unstable winter conditions could also lead to weakened ice and consequently a reduction in ice loads.

Many power plants in Arctic regions have operational restrictions or guidelines during the winter period to avoid ice problems (Foulds, 1988). A shorter ice season will reduce the need to enforce such constraints and permit more optimum use of river flow (Beltaos and Prowse, 2009). By contrast, however, a longer freeze-up period is also likely to increase the period during which such constraints are needed during the remaining ice season.

Some of the most costly ice-induced effects on hydroelectric production result from the blocking of forebays, intakes, and diversion tunnels. Blockage usually results from two sources, both of which could be altered by future ice regimes. First, the intensity and magnitude of frazil ice formation is projected to increase or decrease depending on relative changes in autumn air temperature and flow regimes compared to current climatic conditions (Beltaos and Prowse, 2009). While decreases in frazil ice production will ease constraints on hydropower production, increases (see Section 6.5.1.1) can cause blocking of trash racks and intake structures (Andersson and Andersson, 1992; Ettema et al., 2009), thereby reducing production and increasing operational costs. Moreover, it could also initiate ice problems in downstream river reaches as inflowing production water is forced to bypass intakes. Reaches downstream of intakes are usually characterized by early ice formation and low winter-flow, and sudden releases of water may initiate mechanical break-ups, resulting in ice jamming and erosion damage. In a future with less stable conditions and a longer freeze-up period, this problem may increase in some areas, but it is also likely that it will be reduced in the most southern, temperate river systems. Second, in regions where there is an increase in the intensity or frequency of mid-winter warming spells and, therefore, an increased potential for mechanical winter break-ups (Section 6.5.1), clogging of intakes by drifting ice will cause a loss of water, thus decreasing production. This will be a problem particularly for secondary intakes used in water transfer in high-head systems (Lokna, 2006). Monitoring and mitigation of such problems will be an issue especially for hydropower producers with remote facilities.

Ice jams and subsequent flooding can also threaten hydropower structures in rivers (Beltaos, 2007). Climate change has the capacity both to increase and decrease the magnitude and frequency of these extreme events (Sections 6.4.3 and 6.5.1.1), although their spatial distribution relative to hydropower facilities has not yet been established. In addition to intake and transfer restrictions, ice formation can also influence production through jamming and water level increases in the afterbay.

Although hydropower dams are equipped with spillways to pass floods, the function of which is crucial to dam safety, ice formation can have an impact on the capacity and functionality of these structures (Lia, 1997). In a period with more frequent mid-winter ice break-ups (Section 6.5.1.1), spillway functionality may be affected, particularly in spillway systems with tunnels or gates.

The strength of ice on hydropower impoundments is strongly influenced by reservoir operations such as the lowering of water levels during winter. In a future with shorter winters and a thinner ice cover, particularly along the shoreline and at intakes and outlets, the safety of using reservoir ice for transportation may be compromised. However, such changes in reservoir ice conditions could also lead to some positive impacts for reservoir design and management. At present, a significant amount of reservoir ice is grounded on the banks (e.g., up to 8% of active storage) (Seidou et al., 2007) as water levels are progressively lowered for winter hydroelectric production. Future climate conditions will decrease this volume of inactive storage and reduce some of the current negative consequences, including (i) part of the storage volume being unavailable during winter when electricity demand is high; (ii) grounded ice having the same effect as an additional dead pool storage, forcing the design of larger and costlier structures; (iii) the immobilized water only becoming available at the end of winter when streamflows are large (and demand for electricity is low) thereby increasing flood risk and the probability of spilling; and (iv) grounded ice changing the effective storage curve during winter, which if unaccounted for in dam operations leads to suboptimal decisions (Seidou et al., 2006, 2007). Reductions in any or all of these will provide benefits to hydroelectric operations.

The importance of river ice on hydroelectric operations may also be indirectly affected by future energy adaptations. For example, a reduction in greenhouse gas emissions will require the production of more renewable energy and lead to the introduction of more non-storable energy sources such as windpower. In such a system, load balancing is needed to maintain a continuous supply, and hydropower is ideally suited for this, being renewable, storable in reservoirs, and able to be run with simple start and stop routines (e.g., Benitez et al., 2008). This will have implications for the operational strategies of hydropower producers toward a peaking schedule, and this must be considered when impacts of changes in river ice are evaluated. Generally, peaking operation of hydropower plants in rivers is considered an environmental challenge (e.g., Bradford, 1997; Saltveit et al., 2001; Bruno et al., 2009), and peaking during the ice season further increases potential problems (Scruton et al., 2008). Balancing the variable production from non-storable renewables could lead to a less regular operation of the hydropower system, thereby increasing the potential problems linked to break-up and ice jamming in rivers downstream of hydropower outlets. Additional future variables not addressed here are the potential changes in energy demand and changes in energy prices in the face of future climate. Both of these factors will influence production and thereby have an influence on the future ice problems of the hydroelectric power industry.

6.6. Major uncertainties and future research

6.6.1. Observation networks

- Current river- and lake-ice observation programs in the circumpolar latitudes employ a disparate set of methods and approaches, making it difficult to compare data or conduct large-scale spatial and temporal analyses.

- There should be an international circumpolar effort to assemble and compile a comprehensive freshwater-ice data record that includes data from all forms of available instrumental records.

- There should be a standardization of *in situ* observation methods to facilitate intercomparison of data.

- Given the remoteness of much of the high latitudes, a special focus should be placed on adopting remote sensing approaches to augment the *in situ* networks.

- A number of sites from representative regions should be established around the circumpolar North for conducting long-term monitoring and intercomparison of observational techniques; this will require an international collaborative effort, perhaps one that could be undertaken by an international agency, such as the World Meteorological Organization.

The current state of river- and lake-ice observation in the circumpolar latitudes is composed of a disparate set of programs with varying purposes and approaches. As a result, data about even simple ice phenologies are difficult to compare and observations are often only conducted on regional or national scales, with few attempts to understand change on a circumpolar basis. Although some attempts have been made to centralize the archiving of freshwater-ice information, even these have not been comprehensive of all available records, either because they are difficult to obtain from some countries or regions or because they are difficult to extract from original data sources (e.g., original hydrometric charts and related metadata). In consideration of the above, two recommendations for improvement can be made. The first is that there should be an international circumpolar effort to assemble and compile a comprehensive freshwater-ice data record, which includes data from all forms of available instrumental records and applies relevant quality assurance and control measures. To allow the resulting data to be used in subsequent analyses, there should also be a quantification of data-source errors, such as of temporal error variability for use in time-series analysis of ice phenology dates. Archiving of the original and processed data should be conducted at a recognized international facility, such as the World Data Centre for Glaciology at the National Snow and Ice Data Center, Boulder, Colorado. Second, to improve the comparability of future data, there should be a standardization of *in situ* methods for observing basic

freshwater-ice characteristics. This will also require some form of international collaborative effort, perhaps one that could be undertaken by an international agency, such as the World Meteorological Organization and potentially through their developing program, Global Cryosphere Watch.

Ground-based observations are essential not only to establish time-series evaluations of change, but also to provide essential information for improving understanding of freshwater-ice characteristics, processes, and effects and for validating and improving remote sensing approaches and various forms of numerical simulation models. Where freshwater-ice data are collected as an ancillary product for other purposes, there is an opportunity to improve current observation methods in ways that will vastly improve the value to freshwater-ice studies. For example, many river hydrometric programs measure only part of the total ice thickness when conducting under-ice discharge measurements. A simple modification to the thickness measurement aspect of the program (i.e., complete ice thickness) would make available thousands of additional ice-thickness measurements per year around the circumpolar North. Given the controlling effect of snow, it would also be beneficial to include a simple measurement of surface snow depth. Specific guidelines exist for improving such programs (e.g., Canadian hydrometric surveys, Prowse, 1990; IGOS, 2007) and could be used to guide a circumpolar initiative to revamp and unify circumpolar freshwater-ice *in situ* observing programs. To further broaden at least some forms of ice observation, it would be advantageous to make further use of, or even expand, some of the volunteer networks that have been established, such as Canadian 'Ice Watch' (www.naturewatch.ca/english/icewatch) or the Alaska Lake Ice and Snow Observatory Network (www.gi.alaska.edu/alison).

In unifying the ice observation networks, it is also important that a set of representative regions around the circumpolar North be established for long-term monitoring, preferably including those that already have a lengthy and more comprehensive historical record. Sites with existing paleo-records, or the potential to provide superior forms of such records, might be particularly good candidates. Further to this end, such sites should be the focus of a suite of ice-related modeling efforts, to take advantage of the comprehensive observation datasets, which can be used to facilitate validation and updating of the relevant models. Given this, it would also seem practical in the selection of such 'supersites' to consider the data needs of these models, which could require, for example, information about ecological or socio-economic factors.

Given the spatial scale of the circumpolar North, the remoteness of many regions, and the high costs of *in situ* monitoring programs, remote sensing has long been touted as a promising means to monitor, at least, lake- and river-ice phenology at regional to hemispheric scales. Unfortunately, to date there are only a few remote sensing-derived freshwater-ice operational products, and these do not meet the temporal requirements set by climate programs such as the Global Climate Observing System (accuracy of ± 1 to 2 days for freeze-up and break-up dates) or their spatial resolutions are too coarse (e.g., 4 km IMS product) for ice monitoring on water bodies other than large lakes. Although new approaches have recently been presented in the literature, they require further testing and refinement before application to operational monitoring. Further work should focus on (i) the evaluation of the potential and limits of optical (e.g., Moderate Resolution Imaging Spectroradiometer, MODIS; Advanced Along-Track Scanning Radiometer, AATSR; Medium Resolution Imaging Spectrometer, MERIS) and microwave (e.g., AMSR-E; Advanced Synthetic Aperture Radar, ASAR; and RADARSAT, synthetic aperture radar) data for ice-cover monitoring, and (ii) the development of approaches that make use of optical and microwave data in a synergistic manner for the creation of lake- and river-ice products at a variety of spatial and temporal resolutions for operational and research needs. Key sites need to be established at regional locations around the circumpolar North for validation of all such products. In addition, there needs to be a comparison of conventional *in situ* observations (e.g., ice phenological dates) with historical satellite-derived data (e.g., data from AVHRR; Scanning Multichannel Microwave Radiometer, SMMR; and Special Sensor Microwave Imager, SSM/I). This will ensure some continuity in the transition between the surface-based and satellite observations (e.g., as after the 1980s when *in situ* observations at many lake- and river-ice sites were discontinued). Beyond the observation of basic ice phenology, remote sensing also offers the opportunity to observe some of the dynamics associated with freshwater ice, such as the generation of steep hydraulic gradients and ice-jam floods by river ice during both freeze-up and break-up,

with the use of satellite laser altimeter systems (e.g., GLAS on ICESat or future systems with similar capabilities).

6.6.2. Trend analysis and climatic linkage

- Once a broader freshwater-ice dataset for the circumpolar North is assembled, a comprehensive time series of spatial trends in ice phenologies should be conducted.

- Existing archive data should be further mined for information about changes in temporal and spatial characteristics of river-ice break-up dynamics.

- In exploring the role of climate in controlling trends in freshwater-ice characteristics, assessments should be conducted at a number of scales.

- A special focus should be placed at the scale of major Arctic rivers, including the study of climatic controls exerted in their more temperate mid-latitude headwaters.

With a successful assembly of additional freshwater-ice observational data, there will be an opportunity to more comprehensively analyze time series and spatial trends in ice phenologies and to quantify linkage with similar trends in climatic characteristics and patterns. Even with existing records, however, there is an opportunity to assess more broadly the trends and variability in the temporal and spatial characteristics of river-ice break-up dynamics and to link these to climatic forcing. To date, this has only been conducted for the Mackenzie River in Canada (de Rham et al., 2008b; Goulding et al., 2009b), although similar hydrometric data on which these analyses were conducted exist for other parts of Canada and in other circumpolar countries. Such broader-scale analyses are considered a prerequisite to understanding how climate variability and change has controlled and could control such extreme events.

In exploring the role of climate on a broader regime of trends in freshwater-ice characteristics (i.e., from simple ice phenologies to extreme-event dynamics), assessments should be conducted at a number of scales and approaches. These include basic correlations with simple climatic variables, such as temperature and precipitation, as well as linkage to regional synoptic and large-scale atmospheric circulation patterns, and major atmospheric teleconnections. In the case of large-scale atmospheric patterns, a special focus should be placed on how such systems both control regional variations and produce responses across the circumpolar North. Identification of robust climatic linkage with freshwater-ice characteristics and trends could also offer the opportunity to use such relationships for predicting future effects under altered climatic regimes, although the stability of such relationships under modified climatic conditions also needs to be considered in any such evaluation. Given that some river-ice effects are strongly controlled by catchment-scale processes, such as the dynamics and severity of river-ice break-up fronts, a focus also needs to be placed at the scale of major Arctic river basins, such as the Lena, Mackenzie, Ob, and Yenisey. The river-ice dynamics throughout these river systems are influenced by climatic controls exerted well south of the Arctic Circle in the climatically more temperate mid-latitudes.

6.6.3. Predictive modeling

- To achieve improved prediction of river-ice regimes, advancements need to be made in integrated models that consider future combined changes to landscape hydrology, water-ice-air energy exchanges, in-stream hydraulics, and ice mechanics.

- More validations of existing lake ice models are required across a range of hydro-climatic regimes and lake sizes.

- Predictive modeling of lake- and river-ice systems should be expanded from primarily physical characteristics to include effects on lentic and lotic ecosystems.

- Model validation is a prerequisite for using climate scenarios to evaluate the future effects of climate on ice-controlled aquatic systems, many of which may be characterized by non-linear and/or step changes that are unlikely to be identified fully without such field testing.

Although some prediction of future lake- and river-ice conditions might be possible using climatic analogues (see Section 6.6.2), achieving a more detailed understanding of future freshwater-ice responses suitable for the evaluation of many related impacts will also require the use of physically or mechanistically based models operating at various scales of complexity. In the case of relatively simple lake-ice growth and decay models, achieving a successful explanatory coupling with climatic drivers will provide useful tools for (i) determining what climatic variables are reflected in freeze-up and break-up dates and, hence, how freeze-up and break-up observations might be used as adjuncts to more conventional (e.g., air temperature and precipitation) climate monitoring in data-sparse areas; (ii) understanding sources of uncertainty or complications in interpreting freeze-up and break-up data (e.g., confounding effects of temperature and snowfall); (iii) estimating the magnitude of natural variability in freeze-up and break-up dates for use in climate change detection studies; (iv) estimating the potential impacts of projected climate change on freeze-up and break-up dates, and hence ice-cover duration, as well as ice thickness and type (congelation ice versus white ice); and (v) reconstructing ice phenology and filling in gaps (missing years of observations) in historical records.

In the case of more complex modeling of river-ice conditions, additional cryospheric and hydrological responses to climate will also need to be modeled (see reviews by Prowse et al., 2008; Beltaos and Prowse, 2009). Of particular importance is the role of climate in determining the magnitude of the winter snowpack and its timing and rate of melt, which control the magnitude and rate of the major driving force controlling break-up, jave formation and release, and associated water levels (see Section 6.5.1.1). Similarly, atmospheric conditions control the resisting forces, such as ice thickness and mechanical strength. Achieving suitable coupling of atmospheric and river-ice hydraulic, thermodynamic, and mechanical models will be a major research task, but one that is essential to obtain an understanding of the effects of climate on extreme events in northern systems, make reliable predictions of future conditions, and institute proper adaptation measures (e.g., flood prevention infrastructure; see Section 6.6.4).

Predictive modeling should not be restricted to the purely physical characteristics and effects of lake and river ice. As noted in section 6.5.2, freshwater ice in northern regions is a critical component of aquatic systems and affects a suite of often interrelated physical, chemical, and biological forms and processes. Much of the biological productivity and biodiversity of northern aquatic systems is controlled by ice-related conditions and processes. If the cascading effects of changes in lake- and river-ice are to be understood, then improved response models that include a suite of algorithms linking the interrelated processes need to be produced for northern lentic and lotic systems. While some such models do exist (e.g., Saloranta and Andersen, 2007; see Section 6.4.3), they have not been fully validated or tested under a complete range of hydroclimatic conditions found across the Arctic. Such validation is a prerequisite for using climate scenarios to evaluate the future effects of climate on ice-controlled aquatic systems, many of which may be characterized by non-linear and/or step changes that are unlikely to be identified fully without such testing (e.g., Wrona et al., 2006). As noted in Section 6.6.1, the variability in aquatic systems needs to be factored into the regional selection of long-term representative sites for future observation and physically based studies. Of particular additional importance in site selection should be a consideration of locations that have the potential to be 'hot spots' (i.e., locations that show disproportionately high reaction rates relative to changes in ice and climate conditions) or experience 'hot moments' (i.e., periods of time that exhibit disproportionately high reaction rates relative to longer intervening time periods) (McClain et al., 2003).

6.6.4. Advanced evaluations of socio-economic impacts and adaptation options

- Direct and indirect effects of changes in freshwater ice will both have cascading impacts on socio-economic systems.

- Although some potential adaptation options to such impacts have been identified, they still require economic evaluation and option comparisons.

- Additional cultural values must also be taken into account when considering potential adaptations to traditional ways of life practiced by northerners.

- Only through a broad range of cost-benefit analyses and additional socio-economic modeling can the suitability of relevant adaptation options be properly assessed.

- Given the importance of many ice-affected socio-economic sectors in the Arctic, key locations of such activities should be considered in the selection of long-term observation and validation sites for lake and river ice. By doing so, it is likely to maximize the socio-economic benefits of conducting future freshwater ice and climate change research in the Arctic.

In addition to the effects on aquatic systems, changes in freshwater ice regimes are likely to have cascading effects on socio-economic systems (Section 6.5.3). While it is important to understand how biological systems might adapt to changing ice regimes, it is equally or more important to evaluate how these biological changes might affect socio-economic systems and, more broadly, how other ice-related changes might also affect them. In the first case, it is known that traditional and commercial economies are likely to be affected by changes in aquatic productivity and diversity from altered ice regimes, particularly as they relate to fisheries (Reist et al., 2006a,b,c; Furgal and Prowse, 2008). However, the magnitude of the biological effects remains to be quantified and, similarly, the real cost to traditional and commercial economies. Only through such additional biological and socio-economic modeling of effects can the suitability of relevant, previously defined adaptation options (e.g., changes to fisheries approaches and equipment or relocation of fisheries and possibly resource-based communities) be properly assessed.

In the second case of direct ice-change effects on socio-economic systems, there is also a lack of quantification, primarily because the specific details of expected physical changes to ice regimes remain largely unknown. In Section 6.5.3, some possible adaptation options for hydroelectric facilities (e.g., infrastructure adjustment and changes to operational flow strategies) and ice-road networks (e.g., enhanced ice making, intensification of traffic flow, and land-based road construction) were identified, but these still require economic evaluation and option comparisons. In the case of ice roads, the possibility of adapting to the loss of ice roads through the construction of land-based systems in the Arctic is further complicated by the need to know how related permafrost-affected landscapes will also be affected by changing climate and the costs to construct and maintain such facilities. Just as the responses of lake and river ice to climate change depend on changes in other cryospheric components (e.g., permafrost and snow) so do the adaptation options that need to be assessed through coupled socio-economic models. Furthermore, given the importance of many ice-affected socio-economic sectors in the Arctic, it is recommended that key locations be considered in the selection of long-term observation and validation sites for lake and river ice. By doing so, it is likely to maximize the socio-economic benefits of conducting future freshwater ice and climate change research in the Arctic.

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7. Mountain Glaciers and Ice Caps

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Contents

Key Findings

- Mountain glaciers and ice caps in the Arctic cover an area of around 402 000 km^2 and contain about 0.41 m sea-level equivalent of water.
- Over the past century, nearly all have retreated from maximum extents reached during the Little Ice Age, which ended in the late 19th century. This period of glacier retreat has been associated with an overall reduction in glacier mass during the period of record, which extends to more than 60 years in some cases. Surface mass balance measurements showed negative or nearly balanced conditions and generally showed no trend until the mid-1990s, but since then reveal significantly higher rates of mass loss in Alaska, the Canadian Arctic, and Iceland.
- The fraction of ablation that occurs through iceberg calving can be as much as 40% in regions where it has been measured, but it has not been measured over large areas of the Arctic. Estimation of calving fluxes is therefore a major source of uncertainty in estimates of current and future rates of mass loss from mountain glaciers and ice caps in the Arctic.
- Mass loss (surface mass balance plus calving) from Arctic glaciers probably exceeded 150 Gt/y in the past decade, when it was similar to mass loss from the Greenland Ice Sheet. This suggests that glacier and ice sheet change in the Arctic is probably now the dominant contributor to the eustatic (water mass) component of global sea level rise.
- Under the IPCC A1B emissions scenario, the total volume of Arctic glaciers is projected to decline by between 12% and 32% by 2100 (corresponding to an increase of 49 to 131 mm sea-level equivalent), depending on the choice of general circulation model. These projections are a lower bound since they do not include mass losses by iceberg calving. Regardless, mountain glaciers and ice caps will continue to influence global sea-level changes beyond the 21st century.
- In many parts of the Arctic, climate warming should cause glacier runoff to increase for a few decades or longer, but glacier area reduction will ultimately cause glacier runoff to decline. These changes in glacier runoff will have impacts on water supplies; water quality; hydroelectric power generation; flood hazards; freshwater, estuarine and coastal habitats; and ocean circulation patterns.
- Iceberg hazards to shipping and offshore activities related to exploration for and exploitation of offshore hydrocarbon and mineral resources may increase if changes in tidewater glacier dynamics result in more iceberg production and/or larger bergs, and reductions in sea-ice cover allow icebergs to become more mobile.

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Summary

In addition to the Greenland Ice Sheet, the Arctic contains a diverse array of smaller glaciers ranging from small cirque glaciers to large ice caps with areas up to 20 000 km^2. Together, these glaciers cover an area of more than 400 000 km^2, over half the global area of mountain glaciers and ice caps. Their total volume is sufficient to raise global sea level by an average of about 0.41 m if they were to melt completely.

These glaciers exist in a range of different climatic regimes, from the maritime environments of southern Alaska, Iceland, western Scandinavia, and Svalbard, to the polar desert of the Canadian Arctic. Glaciers in all regions of the Arctic have decreased in area and mass as a result of the warming that has occurred since the 1920s (in two pulses – from the 1920s to the 1940s and since the mid-1980s). A new phase of accelerated mass loss began in the mid-1990s, and has been most marked in Alaska, the Canadian Arctic, and probably Greenland. Current rates of mass loss are estimated to be in the range 150 to 300 Gt/y; comparable to current mass loss rates from the Greenland Ice Sheet. This implies that the Arctic is now the largest regional source of glacier contributions to global sea-level rise.

Most of the current mass loss is probably attributable to a change in surface mass balance (the balance between annual mass addition, primarily by snowfall, and annual mass loss by surface melting and meltwater runoff). Iceberg calving is also a significant source of mass loss in areas such as coastal Alaska, Arctic Canada, Svalbard, and the Russian Arctic. However, neither the current rate of calving loss nor its temporal variability have been well quantified in many regions, so this is a significant source of uncertainty in estimates of the total rate of mass loss. It is, however, clear that the larger Arctic ice caps have similar variability in ice dynamics to that of the Greenland Ice Sheet. That is to say, areas of relatively slow glacier flow (which terminate mainly on land) are separated by faster-flowing outlet glaciers (which terminate mainly in the ocean). Several of these outlet glaciers exhibit surge-type behavior, while others have exhibited substantial velocity changes on seasonal and longer timescales. It is very likely that these changes in ice dynamics affect the rate of mass loss by calving both from individual glaciers and the total ice cover.

Projections of future rates of mass loss from mountain glaciers and ice caps in the Arctic focus primarily on projections of changes in the surface mass balance. Current models are not yet capable of making realistic forecasts of changes in losses by calving. Surface mass balance models are forced with downscaled output from climate models driven by forcing scenarios that make assumptions about the future rate of growth of atmospheric greenhouse gas concentrations. Thus, mass loss projections vary considerably, depending on the forcing scenario used and the climate model from which climate projections are derived. A new study in which a surface mass balance model is driven by output from ten general circulation models (GCMs) forced by the IPCC (Intergovernmental Panel on Climate Change) A1B emissions scenario yields estimates of total mass loss of between 49 and 131 mm sea-level equivalent (SLE) (or 12% to 32% of current glacier volume) by 2100. This implies that there will still be substantial glacier mass in the Arctic in 2100 and that Arctic mountain glaciers and ice caps will continue to influence global sea-level change well into the 22nd century.

As glaciers and ice caps shrink in a warming climate, runoff initially increases in response to higher rates of surface melting. Ultimately, however, runoff will decline as reductions in glacier area outweigh the effect of more rapid melting. This phase of declining runoff does not yet seem to have begun in most regions of the Arctic, but it may begin soon in the Russian Arctic mountains. In the Yukon River basin, Greenland, Iceland, and Norway, glacier runoff is an important resource for hydroelectric power generation, and the viability of hydroelectric projects may ultimately be compromised by runoff decreases associated with glacier shrinkage.

Changes in glacier runoff also result in changes in stream temperature, sediment load, and nutrient export (both magnitude and type) that can be expected to initiate changes in the ecology and productivity of downstream river, lake, and fjord environments. Increased rates of glacier melt may accelerate the release of a range of 'legacy' pollutants stored in firn (partially compacted snow that is the intermediate stage between snow and glacier ice) and glacier ice back into the environment. Increasing freshwater fluxes to fjords and other nearshore marine environments will alter the characteristics of surface water masses and drive changes in circulation. Circulation changes may also follow the retreat of tidewater glaciers onto land, particularly in regions of upwelling close to the termini of these glaciers. These changes can decrease the availability of feeding and resting habitats that are important for marine mammals and seabirds.

The number and size of icebergs produced is likely to change as tidewater glaciers retreat, ultimately reaching zero as their termini emerge onto land. Break-up of floating glacier tongues and ice shelves, a process that has accelerated in the past decade along the northern coast of Ellesmere Island, results in large tabular bergs, while accelerated flow of tidewater glaciers tends to result in accelerated production of small bergs unless flotation of the glacier terminus occurs, when large tabular bergs may be produced. Cessation of small berg production when tidewater termini retreat onto land can reduce the number of such bergs that become grounded in fjords, decreasing the availability of important resting habitat for seals. Circulating icebergs are a potential hazard for shipping, drilling platforms, and seafloor pipelines in the Arctic. Circulation patterns and longevity of bergs may change as the Arctic sea-ice cover declines. Thinner and less extensive sea ice is likely to result in greater berg mobility, while warmer surface waters may result in more rapid melting and disintegration of bergs. A knowledge of the size distributions of bergs produced and circulating in different regions is critical to evaluating the risk of bergs contacting the sea floor and damaging seafloor pipelines.

Glacier retreat will be associated with changes in the magnitude and frequency of a range of geomorphological hazards, most notably outburst floods from ice-marginal and moraine-dammed proglacial lakes, and mass movements from newly deglaciated valley walls. The degree of risk from such phenomena will, however, be highly variable, depending upon the nature of the deglaciated terrain, the size of local populations, and the amount of infrastructure present in individual regions.

Finally, it should be emphasized that the ability to monitor and predict changes in the Arctic's mountain glaciers and ice caps is still quite limited. Basic inventory data for Arctic glaciers are lacking. The number of glaciers on which mass balance is measured is small and declining, and the distribution of measurement sites is highly non-uniform. There are no measurement sites at all in some areas with large glacier areas (such as the Russian Arctic and the Yukon). There is no routine monitoring of mass losses by iceberg calving, and understanding of what controls calving rates is rudimentary. This severely constrains the ability to model calving losses into the future. Studies of the socio-economic impacts of Arctic glacier change are currently few and limited in scope, so most statements about such impacts are based solely on general principles. As such, they do not provide a strong basis for either the formulation or enactment of a policy response to Arctic glacier change. Some of the largest impacts of the ongoing changes in Arctic glaciers (such as global sea-level rise) will be felt in regions of the world that are very far from the Arctic. They may, however, still have social, political, and economic repercussions for Arctic nations – repercussions that need to be explored more thoroughly.

7.1. Introduction

- Mountain glaciers and ice caps cover an area of nearly 402 000 km^2 in the Arctic. Over half of this area is in western North America and the Canadian Arctic.
- The combined volume of these glaciers is sufficient to raise sea level by around 0.41 m if they all melted.
- Glacier types range from small cirque glaciers to large ice caps, with areas of up to 20 000 km^2. These ice caps are dynamically complex, and are drained in part by fast-flowing outlet glaciers that often reach the ocean and lose mass by calving icebergs.
- In many regions, a proportion of the glaciers exhibit 'surge-type' behavior, in which long periods of flow at relatively low speeds are punctuated by short-lived episodes of very rapid flow.
- Long-term changes in the extent, volume, and mass of these glaciers are driven by changes in climate and oceanographic conditions that alter their 'mass balance' – the annual balance between mass gains (due mainly to snowfall) and mass losses (due mainly to surface melting and runoff, and iceberg calving).
- Changes in the thermal structure and flow of glaciers can play an important role in how, and how rapidly, they respond to changing climate and oceanographic conditions.

7.1.1. Background

Glacier ice (including mountain glaciers, ice caps and the Greenland Ice Sheet) occupies about 2.16 million km^2 of the Arctic land surface. This chapter deals with all mountain glaciers and ice caps (hereafter referred to as glaciers), including those in Greenland that are not connected to the ice sheet. The Greenland Ice Sheet itself is addressed in Chapter 8. Mountain glaciers are ice bodies whose geometry and boundaries are controlled by bedrock topography, while ice caps are dome-shaped ice bodies that submerge the underlying bedrock topography. Table 7.1 presents the regional distribution of the

Table 7.1. The areas (rounded to three significant figures) and estimated volumes, in mm sea-level equivalent (SLE), of ice caps and glaciers in the Arctic compared to global values and the ice sheets in Greenland and Antarctica.

Region	Ice-covered area, km²	Volume[a], mm SLE	Source, area
Canadian Arctic	1 51 000	199 ± 30	Ommanney, 1970
Northwestern North America	91 800	71 ± 8	Berthier et al., 2010; ESRI, 2003
Russian Islands[b]	56 700	44 ± 8	Radić and Hock, 2010
Greenland[c]	48 600	44 ± 7	Weng, 1995
Svalbard	36 500	26 ± 2	Hagen et al., 1993
Iceland	11 000	12 ± 6	Björnsson and Pálsson, 2008
Scandinavia	3 100	6 ± 0	Østrem et al., 1973, 1988
Russian Arctic	2 900	4 ± 0	Radić and Hock, 2010
Total (Arctic)	401 600	410 ± 30	
Mountain glaciers and ice caps (global)	741 400	600 ± 70	Radić and Hock, 2010
Greenland Ice Sheet	1 755 600	7500	Bamber et al., 2001
Antarctic ice sheet Antarctic ice shelves	12 348 000 1 555 000	57 000	Fox and Cooper, 1994 (areas). Lythe et al., 2001 (volume)

[a] Volumes of all mountain glaciers and ice caps are from Radić and Hock (2010); [b] Franz Josef Land (13 700 km²), Novaya Zemlya (23 600 km²), Severnaya Zemlya (19 400 km²); [c] Greenland excluding the ice sheet: only the glaciers physically disconnected from the ice sheet are considered here. Areas over 70 000 km² have been reported (Holtzscherer and Bauer, 1954; Weidick and Morris, 1998); however, these estimates have included some glaciers that are connected to the ice sheet but considered independent of the ice sheet in a dynamic sense or that were, by their morphology, discernible as units independent of the ice sheet. Nevertheless, the estimate by Weng (1995) is a minimum estimate because it is based on a 1:2 500 000 map, a scale too coarse to allow identification of many small glaciers.

area and volume of glacier ice in the Arctic. Here, the ice volume is given in units of mm sea-level equivalent (SLE), which is the volume of water (in m³) stored as glacier ice, divided by the surface area of the global ocean (in m²) multiplied by 1000.

7.1.2. Context: What did the Arctic Climate Impact Assessment report say about mountain glaciers and ice caps?

The SWIPA report is an update of knowledge on the state of the Arctic cryosphere relative to the Arctic Climate Impact Assessment (ACIA, 2004, 2005). This dealt with mountain glaciers, ice caps, and the Greenland Ice Sheet in a single section of Chapter 6, '*Cryosphere and Hydrology*'. It gave a regional summary of glacier changes since around 1950 with an emphasis on measurements of surface mass balance and, for regions where mass balance measurements were lacking, changes in glacier area. Mass losses by iceberg calving were mentioned but not discussed in detail. The major ACIA contribution was a series of projections of changes in the surface mass balance of Arctic glaciers for the period until 2100. These mass balance projections were computed using regional seasonal sensitivity characteristics (a measure of the expected change in the annual surface mass balance due to prescribed changes in monthly mean air temperature and precipitation amounts) and projected changes in temperature and precipitation derived from a suite of five GCMs forced by the IPCC B2 emissions scenario. The mass balance projections were used to estimate the potential contributions of different regions of the Arctic to sea level rise over the 21st century.

The ACIA report also commented on the potential impacts of changes in glaciers for other parts of the physical system, ecosystems, and people, and gave an assessment of critical research needs that remains valid today.

7.1.3. Geographic setting

The glacierized areas (i.e. those presently overlain by glaciers) addressed in this chapter are shown in Figure 7.1. Most of the glaciers are located north of 60° N, but some more southerly glaciers, in Kamchatka, Alaska, and northwestern Canada are also included (Table 7.1). The total glacier-covered area is approximately 402 000 km², which is about 54% of all glaciers in the world excluding the large ice sheets (~741 400 km², Radić and Hock, 2010).

Arctic glaciers are irregularly distributed in space (Figure 7.1), and are located in a range of very different climatic regimes (Braithwaite, 2005). In southern Alaska, Iceland, western Scandinavia, and western Kamchatka, the climate is maritime with a relatively small annual temperature range and precipitation rates of a few metres per year, while in the Canadian High Arctic it is very dry, cold and continental, with short summers, a very large annual temperature range (greater than 50 °C), and annual precipitation that ranges from 0.1 to 0.7 m/y. Conditions on Svalbard and in the Russian Arctic islands fall between these two climatic extremes. In northwestern North America, most of the ice cover is in the mountain ranges adjacent to the Gulf of Alaska, while in Arctic Canada the most heavily glacierized regions are in the mountains on Devon and Ellesmere Islands, which are nourished in part by moisture from a large persistent polynya in northern Baffin Bay (the North Open Water).

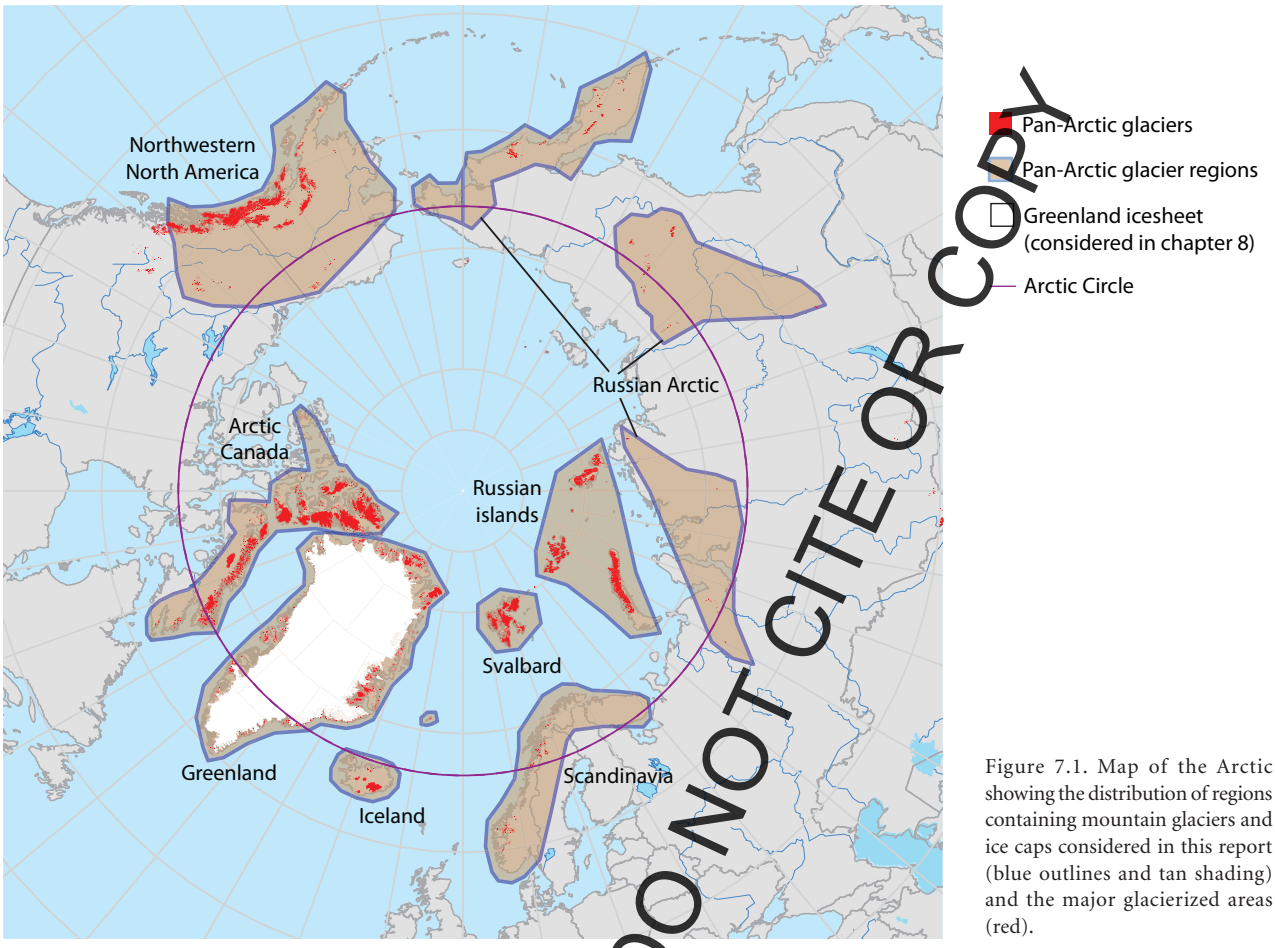


Figure 7.1. Map of the Arctic showing the distribution of regions containing mountain glaciers and ice caps considered in this report (blue outlines and tan shading) and the major glacierized areas (red).

7.1.4. Characteristics of Arctic mountain glaciers and ice caps

7.1.4.1. Morphology

Arctic glaciers have a range of forms. Dome-shaped ice caps have lobes and outlet glaciers that drain ice away from the accumulation area, where annual snowfall exceeds annual surface melt, to lower-lying regions (ablation areas) where the reverse is true, and in some cases to tidewater margins where the ice reaches the ocean and icebergs are calved. Large ice caps are found in the Canadian Arctic, Iceland, Svalbard, the Russian Arctic islands, and in Greenland beyond the margins of the ice sheet (Figure 7.2). In other regions, large glaciers originate from icefields that fill basins within mountain ranges (e.g., in southern Alaska). Many regions, for example Svalbard, also have a large number of individual valley glaciers that occupy valleys and basins in the landscape.

Much of the Arctic ice mass is contained in relatively large ice caps with areas of up to 20 000 km^2, although there are large numbers of independent glaciers with areas ranging from 0.1 to several thousand km^2 (Dowdeswell and Hagen, 2004). The largest ice masses in the Arctic are found in Arctic Canada, and include the Agassiz Ice Cap (17 300 km^2) and Prince of Wales Icefield (19 400 km^2) on Ellesmere Island, the Devon Island Ice Cap (about 14 400 km^2), and the Barnes and Penny Ice Caps (each almost 6000 km^2) on Baffin Island. On Greenland, the largest independent ice cap is the ~9000 km^2 Flade Isblink in the northeast, while the more northerly Hans Tausen Ice Cap, at 3975 km^2, is probably the best studied (Reeh et al., 2001). Austfonna on Nordaustlandet in eastern Svalbard (8120 km^2) is the largest ice cap in the Eurasian Arctic (Dowdeswell, 1986; Hagen et al., 1993). The largest Russian ice cap is the Academy of Sciences Ice Cap (5575 km^2) on Severnaya Zemlya (Dowdeswell et al., 2002), although the northern island of Novaya Zemlya has a larger ice-covered area with many outlet glaciers. In Iceland, Vatnajökull has an area of 8100 km^2 (Björnsson and Pálsson,

Figure 7.2. Small plateau ice caps on Axel Heiberg Island, Arctic Canada, with a larger ice cap in the distance. Note the clear contrast between the snow-covered accumulation area and the greyer ablation area, and the outlet glaciers draining into valleys leading away from the ice cap. Source: Martin Sharp, University of Alberta.

2008). Among the mountain glaciers (Figure 7.3), the largest is Bering Glacier in Alaska, with an area of 3630 km^2.

The larger Arctic ice caps and icefields (Figure 7.4) have complex dynamics involving a mix of fast- and slow-flowing elements that can vary in how they respond to climate changes and variability. Vestfonna on Nordaustlandet in eastern Svalbard is a good example, where four fast-flowing outlet glaciers are embedded within a largely slower-flowing ice cap of 2500 km^2 (Dowdeswell and Collin, 1990). Typical surface velocities may be less than 10 m/y close to the equilibrium-line altitude of valley and cirque glaciers that terminate on land. They can reach many hundreds of metres per year on major calving outlet glaciers, on which seasonal velocity variations can be large, with summer velocities up to an order of magnitude greater than winter velocities (Williamson et al., 2008). Interannual velocity variations can also be significant on such glaciers.

7.1.4.2. Thermal regime

Most glaciers in Iceland, Scandinavia, and central and southern Alaska are predominantly temperate (composed of ice at the pressure melting point). Elsewhere in the Arctic, glaciers tend to have polythermal temperature regimes (composed of a mixture of ice at and below the pressure melting point), and their dynamics may be strongly affected by climate-driven changes in the thermal regime. In some parts of polythermal glaciers, the ice temperature at the glacier bed is below the melting point, implying that the glacier is frozen to its bed. In some smaller (cold) glaciers, all the ice is at temperatures below the melting point, except at the surface where the ice temperature may reach the melting point in summer. The thermal regime of glaciers is determined by the prevailing climatic conditions (snow accumulation and surface energy balance). Climatic change can alter the thermal regime of a glacier, and potentially also its dynamics, because ice deforms more rapidly at higher temperatures and glacier flow can be enhanced by the lubricating effect of meltwater at the glacier bed. Most cold-based and polythermal glaciers are found in dry regions with low accumulation rates. It takes a long time for a climate change signal to penetrate into such glaciers, and changes in the temperature regime are probably not very large on a 100-year timescale. However, in areas where penetration of surface meltwater into cold firn on the glacier surface increases, the release of latent heat when this meltwater refreezes can cause a more rapid change in the thermal regime.

Owing to the polythermal nature of many Arctic glaciers, the formation of superimposed ice by meltwater refreezing on the glacier surface and internal accumulation (where percolating meltwater freezes in cold snow and firn) can be important processes of mass accumulation. When these processes occur, the refrozen water has to be melted again to become meltwater runoff, complicating the measurement and modeling of the surface mass balance. Neglecting or inadequately accounting for these processes in mass balance measurements overestimates mass loss. Superimposed ice formation is important on many Arctic glaciers and is the dominant form of accumulation on some (Koerner, 1970).

Surge-type glacier behavior is common in many parts of the Arctic and can be a source of significant hazards. In a glacier surge, surface velocities increase by an order of magnitude or more and glacier fronts can advance many kilometres (sometimes more than 10 km) in a matter of years. When tidewater glaciers surge, iceberg production increases and can be an important process of mass loss from the glacier (Liestøl, 1973; Dowdeswell, 1989). While the occurrence of individual surges is not directly related to climate change, climate change may alter the frequency of surging (Dowdeswell et al., 1991; Eisen et al., 2001), or even cause glaciers to cease being of the surge type (Dowdeswell et al., 1995). Changes in the number of actively surging glaciers in a region that extend into tidewater can have a large short-term impact on the regional glacier mass balance through changes in the overall calving flux to the ocean.

7.1.4.3. Tidewater glaciers

Iceberg calving plays a major role in the overall mass balance of tidewater glaciers and Arctic ice caps that have significant tidewater margins (Dowdeswell et al., 1997, 2008; Dowdeswell and Hagen, 2004; Błaszczyk et al., 2009) (Figure 7.5). However, accurate calculation of the calving flux requires information that is frequently not available (e.g., ice thickness at the glacier terminus), and modeling of calving flux is challenging because

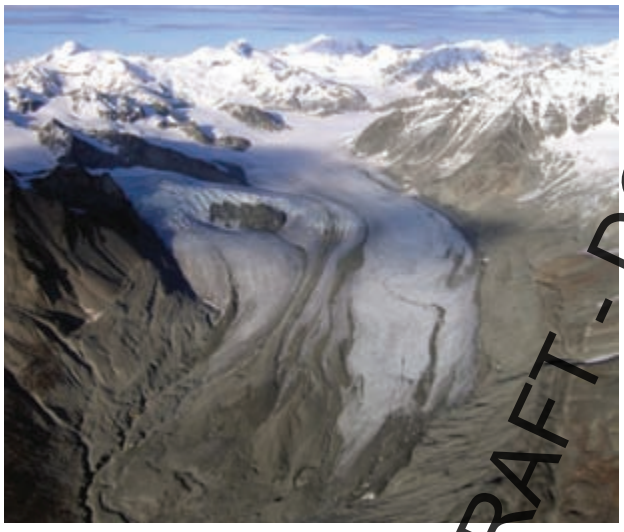


Figure 7.3. Gulkana Glacier in Alaska is a typical mountain glacier and a long-term mass balance monitoring site. Note the end moraines surrounding the glacier terminus that indicate that the glacier was more extensive and thicker in the past. Source: U.S. Geological Survey.

Figure 7.4. The Glacier Bay Icefield, Alaska. Outlet glaciers drain outward from a central accumulation area within the mountains. Source: Landsat imagery courtesy of NASA Goddard Space Flight Center and U.S. Geological Survey.

Figure 7.5. Time-lapse camera overlooking the calving terminus of the tidewater glacier, Kronebreen, in Svalbard. Kronebreen is a focus of study within the IPY GLACIODYN project. Source: Monica Sund, University Centre, Svalbard.

of the lack of a widely accepted calving law. Although individual ice caps may have multiple tidewater glacier outlets, the mass loss by calving is often dominated by one or two of these outlets (Krenke, 1982; Dowdeswell et al., 2002, 2008; Burgess et al., 2005; Williamson et al., 2008; Mair et al., 2009), but calving fluxes from individual glaciers can show large temporal variability.

Recent increases in calving flux from the Greenland Ice Sheet have been associated with acceleration, retreat, and thinning of major outlet glaciers (Howat et al., 2005; Rignot and Kanagaratnam, 2006). Similar changes are observed in Alaska and on some large Arctic ice caps, and may be caused by changes in glacier dynamics related to increased surface melting, penetration of meltwater to glacier beds, and subsequent lubrication of the ice-bed interface allowing increased flow by basal sliding. Other possible causes of calving flux increases include increased melt of the underwater part of terminal ice cliffs due to increasing ocean temperature (Holland et al., 2008), thermal transitions (from cold-based to warm-based) in polythermal glaciers, and break-up and removal of floating ice tongues that formerly restrained the flow of the glacier. Such dynamic changes may be larger and more rapid than those induced by changes in surface mass balance alone.

7.1.4.4. Response to climate change

Glaciers respond to climate change over very different timescales depending on their size, shape, and thermal regime. Among glaciers that terminate on land, smaller glaciers tend to respond more quickly, changing their shape, flow, and terminus position over years or decades. The hypsometry (area-altitude distribution) of a glacier plays an important controlling role in determining its response to changes in climate. In a given region, low-lying glaciers may shrink and retreat quickly at the same time as glaciers at higher elevations grow or maintain their size. Changes in the temperature and salinity of the ocean water adjacent to calving ice fronts can trigger rapid changes in the terminus position, flow velocity, and calving flux of tidewater glaciers, so that large tidewater glaciers may alter their form more rapidly than small glaciers that terminate on land (Holland et al., 2008; Rignot et al., 2010; Straneo et al., 2010). Although Arctic glaciers show a variety of responses to changing climate, it is something to which they are all sensitive. It is worth noting that the surface mass balance of glaciers in the Arctic may be affected indirectly by changes in other components of the Arctic cryosphere, such as the regional snow cover and extent of sea ice, that affect the surface energy balance, atmospheric circulation, and availability of open water as a source of water vapor (e.g., Rennermalm et al., 2009).

7.1.5. Significance and impacts of glacier changes

According to recent estimates, glaciers contributed about 0.8 to 1 mm annually to global sea-level rise during the period 2001 to 2004. The upper limit of the estimates includes the contribution from glaciers surrounding the Greenland and Antarctic ice sheets (Kaser et al., 2006; Solomon et al., 2007). Meier et al. (2007) gave a slightly higher estimate of 1.1 mm for 2006. This is about one third of the total current sea-level rise, or about 60% of that component of the rise attributable to the addition of water mass to the oceans, as opposed to the thermal expansion of ocean waters. The remaining 40% of that component comes from the combined contribution of the Greenland and Antarctic ice sheets. Meier et al. (2007) predicted that a large contribution to sea-level rise over the period to 2100 will still come from glaciers and ice caps. As more than half of the world's glacier and ice cap area is found in the Arctic, it is very important to reduce the uncertainty in estimates of the current and future mass balance of the Arctic glaciers and ice caps.

It is also important to evaluate the impacts of ongoing and future changes in glacier extent and volume on regional water resources, water quality, and the incidence of glacier-related natural hazards. Glacier runoff is exploited as a source of hydroelectric power in Scandinavia, Iceland, Greenland, western Canada, and Alaska. Increasingly negative surface mass balance of the glaciers, and reductions in glacier area may have a direct impact on the water balance of basins used for hydroelectric power generation. Increased freshwater flux from glaciers to nearby fjords and oceans may have an impact on the marine ecosystem via freshening of ocean water and increased transport of nutrients and contaminants from land to sea. Freshening of ocean water may have an impact on fjord circulation and on global ocean thermohaline circulation. Changes in the frequency and magnitude of iceberg calving events may impact infrastructure development, marine transport, fisheries, and oil and gas exploration and production on Arctic continental shelves, where human activity is expected to increase significantly over the coming decades.

Changes in the extent and thickness of Arctic glaciers can destabilize the surrounding terrain and generate geomorphological hazards (rock slides, debris flows, ice avalanches, glacial mudflows, outburst floods), especially when combined with changes in the extent and thickness of permafrost on surrounding slopes. They also create a threat of floods (*jökulhlaups*) from ice-dammed and subglacial lakes, especially in volcanic areas such as Iceland, Alaska, and Kamchatka.

Glacier related tourism is important to some local economies, and shrinking or even disappearing glaciers may have a direct economic impact.

7.1.6. Challenges

The ACIA report (ACIA, 2005) stated that the most difficult task in a regional-scale assessment of glacier behavior is to generalize results from a few glaciers and ice caps to all ice masses in the Arctic. This statement is still valid today. Neither the remote sensing tools nor the glacier and ice sheet models currently available are well suited to studies of regional ice covers that comprise a multitude of glaciers of varying sizes in regions with complex surface topography.

Direct measurements of the mass balance of Arctic glaciers are limited to a small number of glaciers across the Arctic (50 in the 2000 to 2004 pentad); 31 of these were in Scandinavia, which contains only 3.5% of the total mountain glacier and ice cap area in the Arctic. In the past 35 years, the number of *in situ* glacier monitoring sites in the Canadian Arctic and Arctic Russia (regions which contain over 50% of the area of mountain glaciers and ice caps in the Arctic) has declined from 16 in the 1970 to 1975 period, to just six in the period 2000 to 2005, with no measurements in Arctic Russia since 1990 and only one set since 1980 (Dyurgerov and Meier, 2005). There is a pressing need for an updated regional-scale assessment of glacier and ice cap mass balance, the last assessment having been made in 2005. The uncertainties associated with such regional-scale assessments are large owing to the small number of *in situ* measurements and their uneven spatial distribution relative to the distribution of glacier ice.

In situ measurements are critical for quantifying regional mass balance, enhancing process understanding, and validating remote sensing techniques and model predictions. Given the limited number and distribution of such measurements, it is essential to improve and sustain remote sensing capabilities for monitoring ongoing changes in glacier extent, surface elevation and thickness, surface mass balance, ice dynamics, and iceberg production. Although there has been some progress on these issues, there is also a need for repeated, regional-scale mapping of parameters that provide simple indices of glacier mass balance (such as equilibrium-line altitude, glacier facies zones, and summer melt duration; see section 7.4.2) to be used to evaluate year-to-year variability in climate effects on glacier health.

7.2. Climate evolution in glacierized regions of the Arctic

- While some Arctic glaciers have existed throughout the period since the end of the last glaciation, most disappeared during a warm period between 10 000 and 6000 years ago. Many of today's glaciers, therefore, formed in a cooler period after 5000 years ago.
- Although there were several warmer intervals within this cool period, the 20th century appears to have been the warmest century in the past 2000 years, with the warmest conditions occurring between the 1930s and early 1960s and since the mid-1980s.
- Across much of the Arctic, low winter precipitation means that most of the year-to-year variability in glacier mass balance arises from changes in summer temperature. In more maritime regions such as southern Alaska, Iceland, western Scandinavia and Svalbard, variability in winter precipitation can be an important influence on mass balance variability.

7.2.1. Holocene climate

Ice formed during the last glaciation is found in some ice caps in the Canadian Arctic (Koerner and Fisher, 2002; Zdanowicz et al., 2002) and Severnaya Zemlya (Kotlyakov et al., 1991). The absence of such ice in ice caps in Svalbard and other parts of the Russian Arctic is evidence for substantial retreat (or even disappearance) of glaciers and ice caps across much of the Arctic outside Greenland in the early Holocene (Koerner and Fisher, 2002). Therefore, the genesis of much of the present-day Arctic glacier cover may lie in the climate of the middle and late Holocene.

Many climate proxy records suggest that the warmest period of the Holocene in the Arctic was between 10 000 and 6000 years BP, after which the climate cooled, possibly because of a decrease in incident solar radiation at high latitudes due to changes in the Earth's orbital parameters (Koerner and Fisher, 1990; Vinther et al., 2009). Air temperature reconstructions based on ice-core oxygen isotope records from the Agassiz Ice Cap (Canadian Arctic) and Renland (eastern Greenland) suggest peak temperatures slightly in excess of 2 °C warmer than at present during this period (Vinther et al., 2009), which is often referred to as the Holocene Thermal Maximum. However, the exact timing of peak air temperatures during this period seems to have varied across the Arctic. In North America, it was later in the eastern Arctic than in the western Arctic, probably because the residual Laurentide Ice Sheet had a cooling effect on the climate in the east. During the Thermal Maximum, around 7500 years BP, summer temperatures were about 1.6 ± 0.8 °C warmer than in the 20th century (Kaufman et al., 2004). Records of summer melt and oxygen isotope ratios from ice cores in the Canadian and Russian Arctic indicate that cooling following the Thermal Maximum was underway by no later than 6000 years BP (Koerner and Fisher, 2002) (Figure 7.6).

A 2000-year, decadally-resolved, multi-proxy record of Arctic summer surface temperatures (which includes the temperature reconstructions from Agassiz Ice Cap and Renland) shows a cooling trend from 2000 to 100 years BP (Kaufman et al., 2009) (see Fisher, 2002, for an analysis of the reliability of multi-proxy temperature reconstructions). This trend (-0.22 °C/millennium) was probably driven by the orbitally-forced trend in summer insolation at high latitudes over the same period, but may have been amplified in some regions by feedbacks involving the progressive expansion of Arctic sea ice (England et al., 2008; Vare et al., 2009) and seasonal snow cover. The record also shows centennial-scale anomalies about the long-term trend, with the period 450 to 700 AD being cooler than the long-term trend and the period 900 to 1050 being warmer. The coldest period in the record occurred between 1600 and 1850, during the Little Ice Age (Kaufman et al., 2009). There is evidence that this cold period was associated with a persistently negative phase of the North Atlantic Oscillation (Trouet et al., 2009). Although the decline in orbitally-driven summer insolation continued throughout the 20th century, reconstructed Arctic summer temperatures rose sharply, reaching the highest values in the 2000-year record in the period after 1950. On the basis of this reconstruction and recent instrumental data, Kaufman et al. (2009) concluded that the decade 1999 to 2008 had the warmest summers in the 2000-year record (Kaufman et al., 2009).

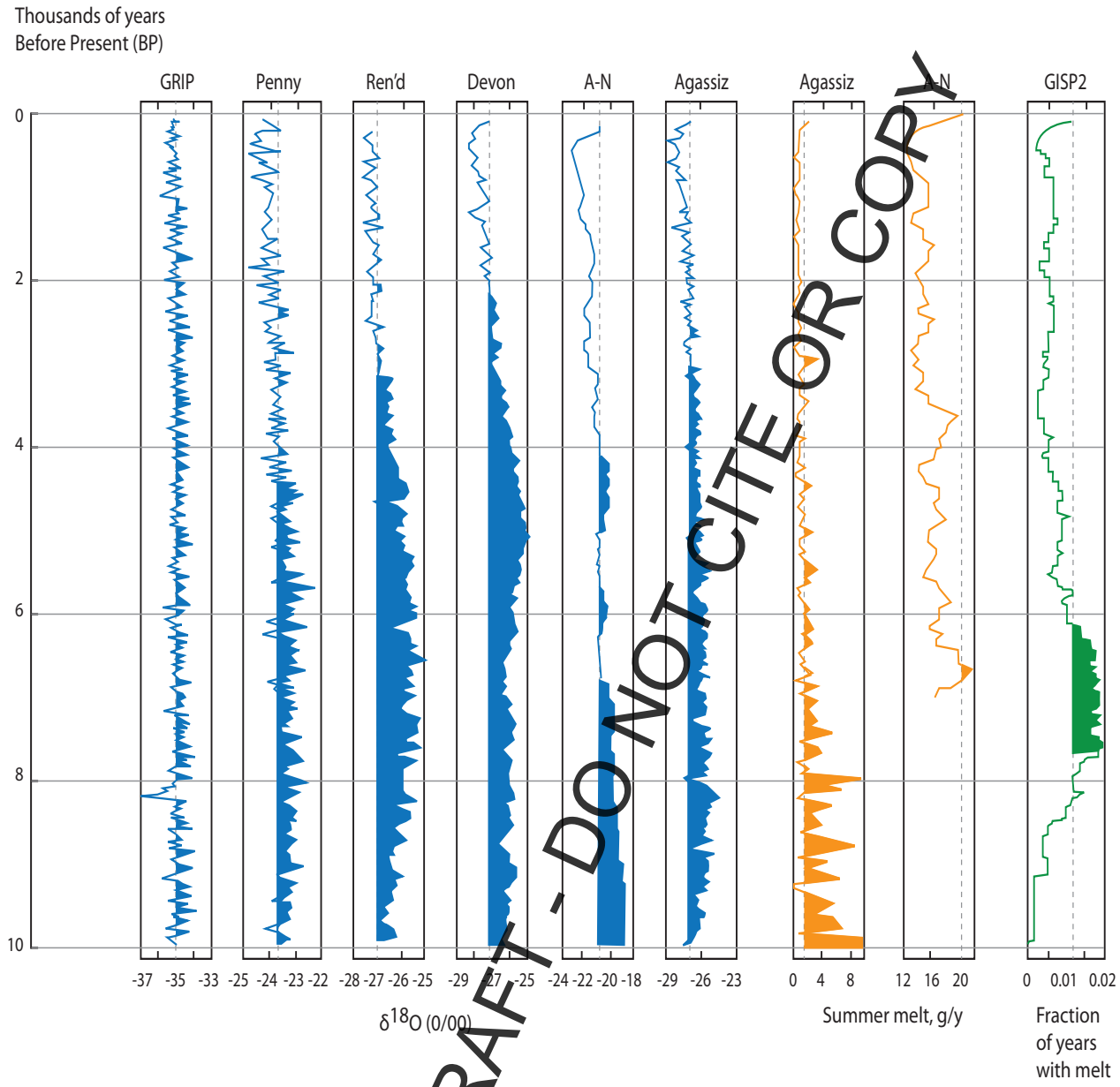


Figure 7.6. Holocene sections of ice cores from Greenland (GRIP, GISP2, Renland), Arctic Canada (Agassiz, Devon, Penny), and Severnaya Zemlya (Akademii Nauk) showing records of δ18O (negative values) and summer melt (positive values), which are proxies for air temperature over the ice caps / ice sheet. Sections shaded in black are sections that suggest conditions warmer than those of today. Source: Koerner and Fisher (2002).

7.2.2. Arctic ice cores: Holocene climate records from glacierized regions

Ice cores from the larger Arctic ice caps yield records of past environmental conditions in glacierized regions that span the full Holocene. The longest records are from the Canadian Arctic, Mount Logan (Yukon), and Severnaya Zemlya (Koerner and Fisher, 1990, 2002; Kotlyakov et al., 1991; Fisher et al., 1995, 2008; Koerner, 1997; Zdanowicz et al., 2002; Kinnard et al., 2008). The melt layer record from the Agassiz Ice Cap and the oxygen isotope record from Severnaya Zemlya show a temperature maximum in the early Holocene, before 8000 BP, while oxygen isotope records from Arctic Canada typically show a later thermal maximum – as late as 5000 BP on Devon Island (Paterson et al., 1977). Melt layer records show a persistent cooling trend from about 8000 BP until the mid-19th century, after which there is renewed warming. Oxygen isotope records also show cooling, but the timing of this is delayed until after about 5000 BP in some records (Koerner and Fisher, 2002). The melt records suggest maximum summer cooling from the Holocene Thermal Maximum of 2.0 to 2.5 °C, while the oxygen isotope records suggest cooling of 1.3 to 3.5 °C (Koerner and Fisher, 2002). Part of the early Holocene warming recorded in the ice core records must be due to the effects of surface lowering due to ice cap thinning, although this would have been partly offset by cooling induced by isostatic uplift of the land surface as the ice cover thinned and shrank (Koerner and Fisher, 2002; Vinther et al., 2009).

Oxygen isotope records from ice cores from smaller ice caps in Arctic Canada, northern Greenland, Svalbard, and the Russian Arctic islands show temporal variability but not the long-term cooling trend apparent in the records from the larger ice caps. This suggests that the small ice caps began to re-grow

in the latter part of the Holocene after melting away during the thermal maximum (Hammer et al., 2001; Landvik et al., 2001; Koerner and Fisher, 2002). The record from the upper 125 m of the ice core from Hans Tausen Iskappe in northern Greenland suggests a warm period between 900 and 1100 AD, and that the period 1700 to 1900 was the coldest in the past 2000 years. There was strong warming from the 1920s to the early 1960s (making the 20th century the warmest part of the record), but no clear trend in temperatures thereafter (Hammer et al., 2001). New oxygen isotope records from Austfonna and Lomonosovfonna in Svalbard span the past 600 to 800 years and show gradual cooling to a minimum in the 19th century, followed by rapid warming around 1900 that made the 20th century the warmest century in the past 600 years (Isaksson et al., 2003, 2005). A 275-year oxygen isotope record from Akademii Nauk Ice Cap on Severnaya Zemlya shows the coldest conditions around 1790, and then continuous warming until 1935. It also shows that the 20th century was the warmest period in this region (Fritzsche et al., 2005). A reconstruction of snow accumulation rates on Lomonosovfonna shows an increase of 25% in the late 1940s (Pohjola et al., 2002). A similar increase is recorded on Severnaya Zemlya after 1935 (Opel et al., 2009).

The oxygen isotope record from Mount Logan is unusual in that it appears to be a proxy for changes in the moisture source region for precipitation, rather than air temperature (Fisher et al., 2008). Major changes in moisture source region seem to be associated with switches between strong, frequent El Niño conditions (associated with meridional patterns of atmospheric flow and water vapor transport towards the Yukon) and strong, frequent La Niña conditions (associated with more zonal flow and moisture transport towards the Pacific Northwest). La Niña periods are associated with reduced precipitation (and snow accumulation) in the southern Yukon, while the reverse is true for El Niño periods. Periods of enhanced meridional flow seem to have occurred around 4200 BP and between 8000 and 7000 BP, while the modern El Niño Southern Oscillation regime seems to have been initiated after 4200 BP (Fisher et al., 2008).

Chemically-based melt indices from the Lomonosovfonna ice core show very high melt in the 12th century (Grinsted et al., 2006), as do melt layer records from the Canadian Arctic. This suggests a warm episode in medieval times in these parts of the Arctic. This warm episode was followed by a long period of reduced melt lasting to the mid- to late 19th century, after which melt increased sharply. For example, the melt layer record from Akademii Nauk Ice Cap shows maxima in the 1840s, 1880s, and from 1900 to 1970, after which melt layer content dropped sharply (Opel et al., 2009). However, the increase in melt was delayed until about 1925 in the new record from Prince of Wales Icefield, Ellesmere Island (Kinnard et al., 2008). The first principal component of seven Arctic melt layer records spanning the period 1551 to 1956 explains 34% of the variance in these records. It shows a strong increase in melt layers starting around 1830 and peaking in the mid-20th century (Kinnard et al., 2008) and has a similar form to a multi-proxy summer temperature reconstruction for the Arctic (Overpeck et al., 1997). Melt layer records from multiple sites on the Devon Island Ice Cap show an increase in ice fraction of over 50% after 1989 relative to the period 1963 to 1988 (Colgan and Sharp, 2008), consistent with a shift to increasingly negative surface mass balance after 1987 (Gardner and Sharp, 2007).

7.2.3. Climate records for glacierized regions of the Arctic derived from instrumental observations and climate re-analysis

Observational records of past surface air temperature from around 70 stations in the Arctic (north of 62° N) since 1875 show large multi-decadal fluctuations. Maxima in mean annual air temperatures occurred in the 1930s and 1940s and in the past two decades, with minima before 1920 and from about 1955 to 1985 (Polyakov et al., 2003). The same trends are apparent for the region 68° to 90° N in the 250-km smoothed air temperature dataset from the Goddard Institute for Space Studies. Mean summer 2-m air temperature anomalies in this dataset show a similar history to mean annual anomalies, but the range of summer anomalies is smaller than for the mean annual air temperature. In addition, there is a period of high mean summer air temperature in the 1950s that is not apparent in the record of mean annual air temperature (Figure 7.7).

Correlations between surface air temperature measured at coastal weather stations and the monthly North Atlantic Oscillation index are positive for a region that includes northern Scandinavia, Svalbard, and the Eurasian Arctic islands, and negative for a region that covers most of Greenland and the Canadian Arctic (Polyakov et al., 2003). Since the late 1990s, surface air temperature anomalies over the Arctic Ocean have become increasingly positive in autumn, especially over those regions where there has been a strong decrease in September sea-ice extent (Serreze et al., 2009). This trend may have resulted in longer melt seasons over glacierized regions of the Canadian and Eurasian Arctic.

In most regions of the Arctic, low winter precipitation means that interannual variability in annual glacier mass balance arises from variability in the summer balance, which is strongly controlled by summer air temperature. Gardner et al. (2009) found positive correlations between mean lower-troposphere summer air temperature from climate re-analyses and surface air temperature measured at the summit elevations of Arctic ice caps. Sharp and Wang (2009) reported similar

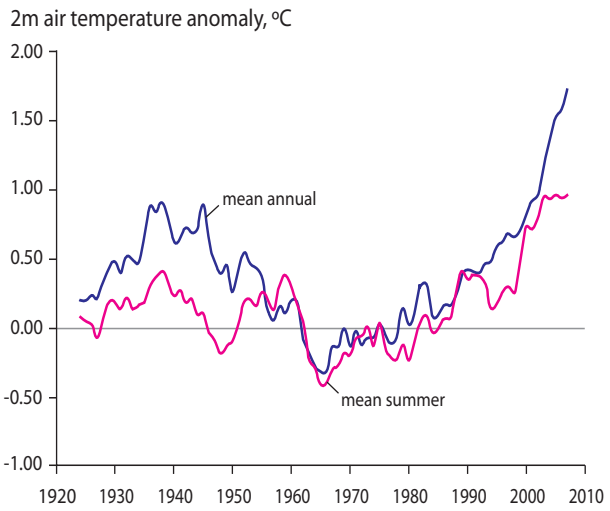


Figure 7.7. Five-year running means for mean annual and mean summer (June – August) 2-m air temperature anomalies (relative to the mean for 1951 to 1980) for the region 68° to 90° N from the Goddard Institute for Space Studies 250-km smoothed dataset.

correlations between lower tropospheric summer mean air temperature from climate re-analyses and melt season duration on Eurasian Arctic ice caps. For the period 1948 to 2008, NCEP/NCAR (U.S. National Weather Service National Centers for Environmental Prediction and the National Center for Atmospheric Research) Reanalysis (Kalnay et al., 1996) decadal mean 700 hPa air temperatures over glacierized regions of the Arctic in June to August were warmest in either the 1950s (Russian Arctic, northern Ellesmere and Axel Heiberg Islands) or the 2000s (Alaska, Iceland, Svalbard, and the rest of the Canadian Arctic). They were coldest in the 1990s in the Eurasian Arctic and the 1960s and 1970s over Iceland, Alaska, and the Canadian Arctic.

In the more maritime regions of the Arctic (such as Alaska, Iceland, Svalbard, and western Scandinavia), high variability in winter accumulation can also induce interannual variability in the annual mass balance. According to the NCEP/NCAR Reanalysis, decadal mean winter (September to May) precipitation anomalies were most negative in the 1950s over Alaska, the 1960s over Iceland, and the 1980s over Svalbard. They were most positive in the 2000s over Alaska and the 1950s over Iceland and Svalbard.

High summer precipitation keeps glacier surface albedo relatively high and reduces summer melt, while low summer precipitation has the opposite effect. In the Eurasian Arctic islands, decadal mean summer precipitation anomalies were typically positive from the 1950s through the 1970s and negative from the 1980s to 2000s. The opposite trend was observed in Iceland.

7.3. Changes in glacier extent, volume and total mass

- Following the early Holocene warm period, the cooler climate between 6000 years BP and the end of the Little Ice Age resulted in glacier growth and advances across the Arctic.
- In some cases, maximum glacier extents were reached as early as the mid-17th century, but in most regions glaciers were at or close to their maximum extents in the late 19th and/or early 20th centuries.
- Across the Arctic, glaciers began retreating and losing mass in the early 20th century. This trend continued through the mid-20th century, although glaciers in some regions experienced brief episodes of slower retreat, reduced mass loss, or even mass gain. Overall, there were substantial reductions in glacier area and mass across the entire Arctic over the 20th century.
- Mass loss has accelerated across most regions of the Arctic since the mid-1990s.
- In some regions, the timing and duration of episodes of relatively positive or negative mass balance were tied to major changes in atmospheric circulation related to phenomena such as the North Atlantic Oscillation (Iceland, western Scandinavia), the Pacific Decadal Oscillation (western Canada, southern Alaska), and the location and intensity of the summer circumpolar vortex (Arctic Canada).

7.3.1. Arctic glacier changes during the Holocene

Reconstructions of dated lateral and terminal moraine positions (and associated stratigraphy) are the primary source of information on glacier changes during the Holocene. In western North America, the Cordilleran ice sheet reached its maximum extent and thickness by 16 000 BP. Between 15 000 and 11 000 BP, climate conditions allowed some lobes of the ice sheet to advance (Menounos et al., 2009), and alpine glaciers also advanced in the southern and north-central ranges of the Cordillera, but by 11 000 BP glacier extent was comparable to that at present (Clague et al., 1982). Glacier extent reached a minimum between about 11 500 to 9000 BP, with temperatures 2 to 3 °C above present and generally drier conditions (Kaufman et al., 2004). Subsequently, periods of glacier advance occurred from 8600 to 8200 BP, 7400 to 6500 BP, 4400 to 4000 BP, 3500 to 2800 BP, and 1700 to 1300 BP (Menounos et al., 2009), and maximum Holocene extents were reached during the early 18th or mid-19th centuries AD.

A number of ice caps in Arctic Canada (Devon, Agassiz, Barnes, Penny) contain ice dating from the last glaciation (Paterson et al., 1977; Hooke and Clausen, 1982; Fisher et al., 1983, 1995, 1998; Zdanowicz et al., 2002) and must therefore have survived throughout the Holocene. During deglaciation, the margins of the last Innuitian ice sheet covering the Queen Elizabeth Islands reached the current margins of ice caps on Devon, Ellesmere, and Axel Heiberg Islands between 10 000 and 7500 ^{14}C years BP (England et al., 2006). Many of the smaller ice caps and glaciers in this region disappeared during this period (Koerner and Fisher, 2002). Glaciers began to grow again after 1000 ^{14}C years BP (Blake, 1981), reaching maximum extents in the late 19th or early 20th centuries. On Baffin Island, the Laurentide Ice Sheet retreated progressively throughout the Holocene to its current margin, that of the Barnes Ice Cap. Deglaciation was especially rapid around 7000 BP, when the ice cover in Foxe Basin collapsed (Briner et al., 2009). The early Holocene ice extent in this region is poorly constrained but glaciers began to advance again as early as 6000 BP, reaching extents similar to those of the Little Ice Age (the Holocene maximum for alpine glaciers on Baffin Island) by 3500 to 2500 BP (Briner et al., 2009). Small ice caps have been present in the northern interior of Baffin Island for 1600 to 2800 years, but they were absent for much of the middle Holocene. These ice caps started to expand after 1280 AD, and especially after 1450 (Anderson et al., 2008).

Most glaciers in Greenland were smaller during the early Holocene than at present and some areas that are currently occupied by mountain glaciers and ice caps may have become completely ice free in that period (Hammer et al., 2001). Neoglacial moraines have been reported (from western and southeastern Greenland), but in most regions historical advances were the most extensive after the early Holocene. In all areas apart from northern Greenland (where many ice margins are stationary), local glaciers are currently receding from their historical maximum extents (Kelly and Lowell, 2009).

Iceland was largely ice free during the early Holocene (8000 to 6000 BP), after which Neoglacial cooling began. Glacier extent increased between 4500 and 4000 BP, and again between 3000 and 2500 BP. The most extensive Holocene ice extent was

reached during the period 1250 to 1900 AD – probably in the latter part of that period between 1700 and 1850 (Geirsdóttir et al., 2009). However, some steep alpine glaciers reached their maximum extents earlier, probably around 2500 BP, when the present ice caps are also likely to have formed. A general glacier recession that began in the 1890s accelerated after 1930. Retreat slowed by the 1960s as a result of cooler summers beginning in the 1940s, and many steep glaciers were advancing in the 1970s. Renewed climate warming since 1985 has led to a new episode of retreat, and all major non-surging outlet glaciers have been retreating since 1995, at rates of up to 100 m/y. The southern outlets of Vatnajökull eroded down into soft sediments during the pre-1890 advance and have been especially susceptible to rapid retreat since then. The volume of Vatnajökull has decreased by about 300 km^3 (10%) since 1890 (Björnsson and Pálsson, 2008).

In northern Scandinavia, retreat of the Scandinavian Ice Sheet during the Late Glacial and early Holocene was punctuated by numerous readvances in the period 11 200 to 8000 BP. Most glaciers disappeared completely at some point in the early to mid-Holocene, and ice cover is likely to have reached a minimum between 6600 and 6300 BP (Nesje, 2009). Subsequently, there was a period of renewed glacier growth during which there were numerous century- to millennium-long 'Neoglacial' events. Glaciers in northern Sweden were probably at their Little Ice Age maximum extent in the 17th and early 18th centuries (Karlén, 1988), while those in Norway reached their maxima in the mid-18th century (Winkler, 2003; Nesje, 2009), or even later in some parts of northern Norway (Bakke et al., 2005). Most Swedish glaciers reached an extent close to their Holocene maximum at the beginning of the 20th century (Holmlund, 1993). Since then, Scandinavian glaciers have generally retreated, particularly since the 1930s (Nesje, 2009). Increased snow accumulation caused some glaciers to advance in the 1990s (Dowdeswell et al., 1997).

In northeast Svalbard, many glaciers started to retreat after the Last Glacial Maximum by about 9500 BP (Blake, 1989). During this period of deglaciation, variations of the glaciers and ice caps in Svalbard were similar to those of glaciers in Scandinavia. However, the Younger Dryas advance was apparently small due to a very dry climate, and the Little Ice Age maximum extent was larger than the Younger Dryas advance (Mangerud and Landvik, 2007). In the early Holocene, glaciers were much smaller than at present and perhaps even completely absent. A period of glacier growth from 3000 to 2500 BP was followed by a warmer period with smaller glaciers (Svendsen and Mangerud, 1997). Dating of relict vegetation found under the cold-based Longyearbreen indicates that late Holocene glacier growth started about 1100 BP. This period of glacier growth culminated in the Little Ice Age maximum with the greatest Holocene glacier extent. Many ice-cored moraines were formed at this time, which may have been as recently as around 1900 (Humlum et al., 2005). Surging of Svalbard glaciers is common and the maximum extent of many glaciers may be a result of surge dynamics (Hagen et al., 1993). Since the Little Ice Age maximum, the glaciers have generally retreated in response to early 20th century warming (Lefauconnier and Hagen, 1991; Hagen et al., 2003a,b).

In the Russian Arctic, outlet glaciers in northwestern Novaya Zemlya retreated from the outer coast by the earliest Holocene. Prior to about 9500 BP the terminus of the tidewater Shokal'sky Glacier was over 1 km behind its present margin, allowing incursion of the sea (Zeeberg, 2001). A sediment core from Russkaya Gavan spans a period of 800 years and allows reconstruction of the recent history of Shokal'sky Glacier. A noticeable advance around 1400 AD was followed by a major retreat by about 1600. Increased melting in the 1900s is suggested by an increase in sedimentation rates (Polyak et al., 2004). On Franz Josef Land, glaciers were less extensive than at present throughout the period 10 300 to 4400 ^{14}C years BP, but re-expanded to their present margins by 2000 ^{14}C years BP (Lubinski et al., 1999). A major advance occurred during the past 1000 years, with maxima around 1200, 1400, and in the mid-17th century. Glaciers were more extensive than now at the start of the 20th century; since then there has been widespread glacier retreat.

7.3.2. Arctic glacier changes in the 20th and 21st centuries

7.3.2.1. Methods for determining glacier mass balance

Mass balance is the change in mass of a glacier, quantified as the difference between mass gains (accumulation) and mass losses (ablation), over a specified time period. A summary of the processes that determine the mass balance of a glacier is provided in Appendix 7.1, and depicted in Figure 7.8 (see also Box 7.1). Over the 20th and beginning of the 21st centuries, many new methods have been developed for measuring glacier mass balance. Each method provides information at unique spatial and temporal resolutions, and samples specific parameters to characterize glacier evolution over time. Important challenges for global mass balance inventories include the integration of these different datasets and quantification of the errors in each approach.

Long-term records of surface mass balance that form the basis of existing global datasets have been acquired from direct

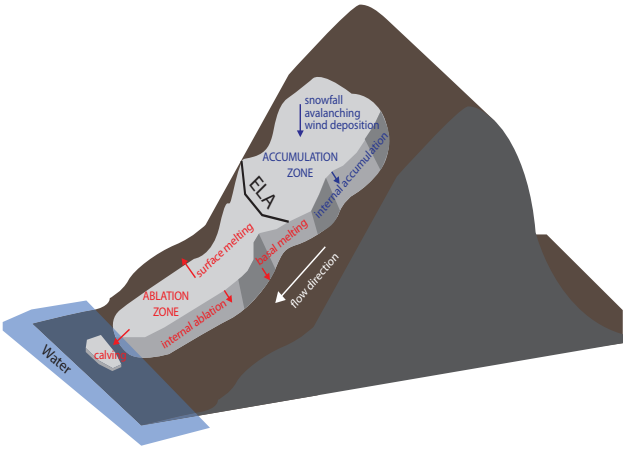


Figure 7.8. Terminology relevant to the mass balance of a mountain glacier or ice cap. Arrow direction indicates whether mass is being added or removed. The color of the arrow indicates the phase in which the addition or removal of mass occurs. See the definition of 'mass balance' in Appendix 7.1 for further explanation. Source: Regine Hock (University of Alaska), E. Leinberger (University of British Columbia), and A. Arendt (University of Alaska); modified from Cogley et al., 2011).

Box 7.1. Units for mass balance measurements

Mass balance measurements have units of mass per unit time (see Appendix 7.1). It is also common to report the rate of change of mass per unit area, for example, kg/m2 per year, which is numerically equivalent to the millimetres of water equivalent (w.e.) per year. In this report, mass balances are usually presented in units of gigatonnes per year (Gt/y) or metres of water equivalent per year (w.e./y). Exceptions are those studies reporting cumulative mass balances, in which case they are not converted to a rate because it is not always clear what exact time period was used for the measurements, and those studies reporting volume changes, in which case they are not converted to a mass equivalent unless the density of the changing snow and ice volume is clearly stated.

sampling of individual glaciers using glaciological methods (Østrem and Brugman, 1991; Cogley, 2005). In this approach, measurements of snow accumulation are acquired from snow probing or digging snow pits to measure snow depth and density. Glacier ablation is measured by determining the vertical change in distance from the glacier surface to a marker, fixed within the reference frame of the moving ice mass, and by assuming an appropriate density for the snow, firn, or glacier ice that melted. Annual mass balance is defined as the difference between the annual accumulation and the annual ablation. Point measurements of this quantity are extrapolated to the entire glacier. Two processes, internal accumulation and the formation of superimposed ice, involve the refreezing of some fraction of the annual melt and alter surface mass balance by reducing the amount of meltwater runoff. However, not all surface mass balance programs measure these processes owing to the difficulty of doing so, resulting in overestimates of the rate of mass loss. Regardless, glaciological mass balance measurements generally provide the most accurate sampling of mass balance at a point, but are limited in spatial extent.

Developments in ground surveying and remote sensing technologies have resulted in many geodetic mass balance methods. These calculate a volume balance determined from repeat measurements of surface elevation and area, from which the mass balance is obtained by estimating the density of the changing ice volume. Elevations are determined using overlapping aircraft or satellite imagery to determine the 3-dimensional position of the surface through stereoscopy, ground mapping of the glacier surface elevation using theodolite or Global Positioning System (GPS) surveying, or airborne / satellite altimetry using radar or laser ranging instruments. Geodetic methods provide a broad regional sampling of glacier changes, but can have large errors due to uncertainties in the density of the volume lost, especially over short periods, non-representative sampling of fast- and slow-flowing sectors of ice caps, or biases in the distribution of surface-elevation change measurements relative to the distribution of rates of elevation change (e.g., in aircraft altimetry where measurements are typically made along a glacier centerline where rates of thickness change may be larger than the glacier-wide mean; Berthier et al., 2010).

Recent advances in spaceborne geodesy provide gravimetric data to determine glacier mass variations via measurement of changes in Earth's gravity field. At present, this is achieved using the Gravity Recovery and Climate Experiment (GRACE) satellites that measure variations in Earth's gravitational potential. Corrections must be made to the satellite measurements to account for all non-glacier sources of mass variation, such as glacial isostatic adjustment and changes in terrestrial water, atmospheric and ocean mass. Gravimetric methods circumvent the need for density corrections of measured volume changes, but are limited in spatial resolution (about 50 000 km^2) and can only effectively sample broad regions undergoing significant change.

Measurements of glacier thickness are essential for estimating glacier volume and for computing mass losses by iceberg calving. They are also needed as input to models of the dynamic response of glaciers to climate forcing. They are usually obtained using the technique of radio echo sounding from either ground-based or airborne platforms (Dowdeswell and Evans, 2004). Airborne campaigns are typically required for measurements of the thickness of larger glaciers and ice caps, as well as for regional-scale studies (Dowdeswell et al., 2002, 2004).

Area/volume scaling (Bahr et al., 1997) is used to estimate glacier volume when glacier areas are known but direct measurements of volume are lacking. This method is based on the observed power-law relationship between a glacier's volume and area. Glacier surface areas measured at different times are used to estimate corresponding glacier volumes, from which changes in glacier volume can be determined. This method can have large errors when data are not available to constrain the scaling relationship and determine suitable scaling coefficients (Radić et al., 2008). Therefore, it is most appropriately applied to entire populations of glaciers rather than to individual ice masses.

Changes in proxy indicators are also used to approximate mass balance trends, and these approaches are summarized in Section 7.4. The distribution and length of glaciological and geodetic mass balance records in the Arctic region are highly variable, but the longest continuous records (from Scandinavia) extend back to the 1940s (Figure 7.9). Records from individual regions are discussed in the following sections.

7.3.2.2. Northwestern North America

Northwestern North America includes all glaciers of Alaska (United States) and Yukon Territory (Canada) and those glaciers in British Columbia and Alberta (Canada) that drain into the Alaska Panhandle or into the Arctic Ocean via the northern tributaries of the Mackenzie River.

During the 20th century most land-terminating glaciers in northwestern North America retreated extensively from their Little Ice Age maximum extent (Kaufman et al., 2004). Most of the recession occurred between 1900 and 1950 and from 1980 to present, with a period of slower recession, and in some cases advance, between 1950 and 1980. Since 1980, nearly all glaciers have been in a state of retreat. For example, in British Columbia, the extent of glaciers in the northern Coast and St. Elias mountains changed by -7.7 ± 2.9% and -7.9 ± 2.6%, respectively, over the period 1985 to 2005 (Bolch et al., 2010). Between 1958/1960 and 2006/2008, the total glacier area in the Yukon Territory decreased by 2541 ± 189 km^2 (or 22% of the

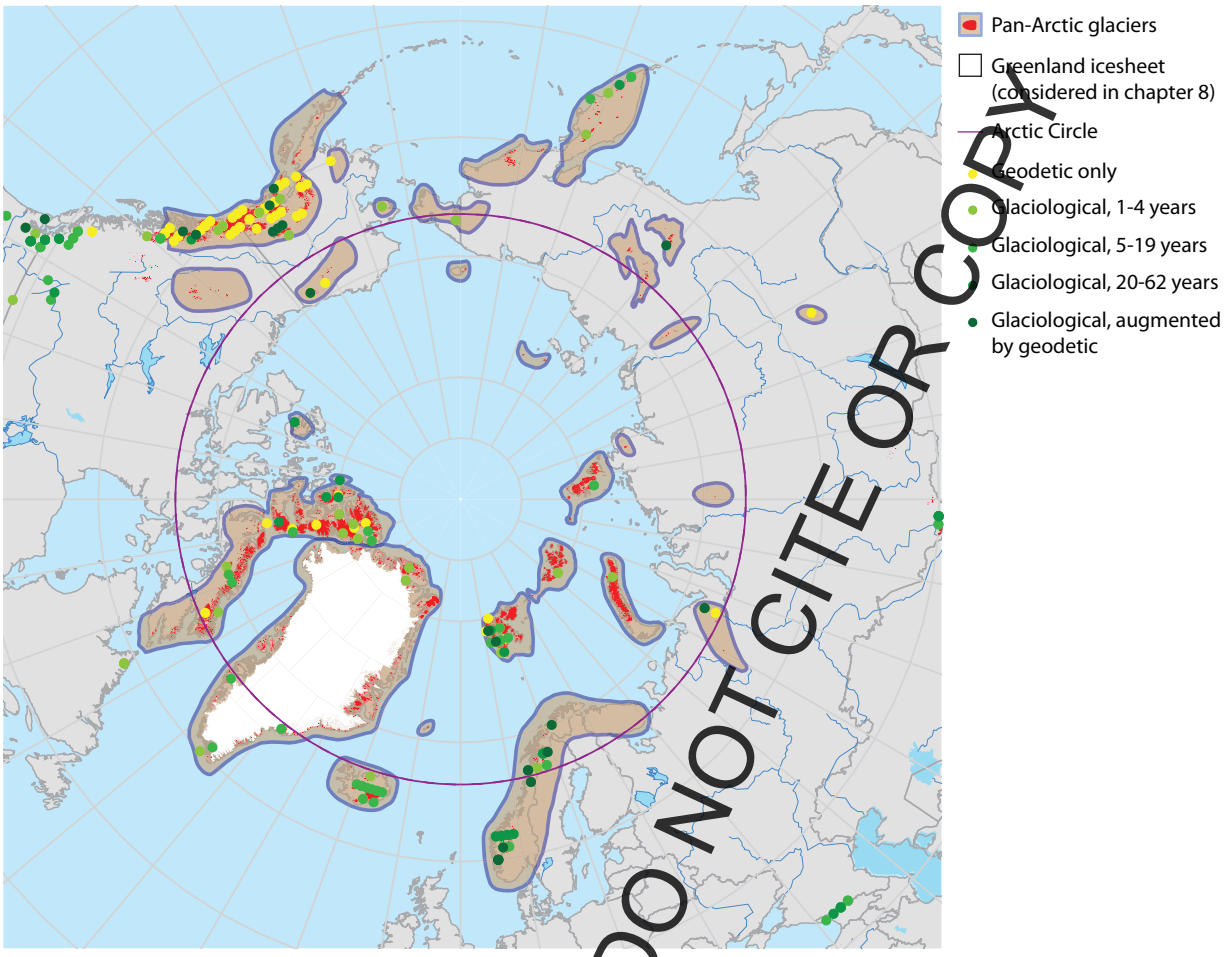


Figure 7.9. Glacierized regions in the Arctic and the locations where mass balance measurements have been made. The length of time of the geodetic and glaciological mass balance records are also indicated. Source: AMAP.

initial area of 11 622 km^2) (Barrand and Sharp, 2010).

Regular glaciological mass balance measurements in Alaska began during the 1960s and trends generally agree with observed variations in glacier extent. Wolverine Glacier lost mass from 1966 to 1976 and gained mass from 1976 to 1988, while Gulkana Glacier lost mass at a constant rate from 1966 to 1973 and at an increasing rate from 1974 to 1985 (Josberger et al., 2007). After about 1988, both glaciers entered a period of accelerating mass loss (Figure 7.10).

Aircraft laser altimetry measurements showed that about 85% of 67 glaciers in northwestern North America had a mass balance of -52 ± 15 Gt/y between 1950/1970 and the early 1990s (Arendt et al., 2002; Echelmeyer et al., 1996). A repeat sampling of 28 of these glaciers between the early 1990s and late 1990s / early 2000s indicated a doubling of the rate of mass loss (-96 ± 35 Gt/y). A recent analysis compared digital elevation models (DEMs) from the 1950s/1970s against those from the 2000s, and argued that estimates based on centerline aircraft altimetry overestimated the mass loss of Alaskan glaciers by 34% (Berthier et al., 2010). These new estimates suggest a mass loss rate of 41.9 ± 8.6 Gt/y for the period 1962 to 2006.

Other geodetic measurements have been acquired through differencing of 1950s U.S. Geological Survey (USGS) and NASA's Shuttle Radar Topography Mission (SRTM) DEMs. Glacier mass balances in regions south of 60.5° N in southeastern Alaska and the Kenai Mountains were -15 ± 4.0 Gt/y (Larsen et al., 2007) and -1.2 ± 0.25 Gt/y (VanLooy et al., 2006), respectively, during a 30- to 50-year period ending in 2000. Another study using the SRTM data found the glaciers of the northern Coast and St. Elias mountains had a mass balance of -11.1 ± 2.0 Gt/y during the period 1985 to 2000 (Schiefer et al., 2007). Recent altimetry measurements for 2004 to 2008 suggest an acceleration of mass loss relative to the 1990s, especially for glaciers at low elevations (Lingle et al., 2008).

There are three estimates for the mass balance of glaciers surrounding the Gulf of Alaska based on data from GRACE: -110 ± 30 Gt/y for the period 2002 to 2004 (Tamisiea et al., 2005); -101 ± 22 Gt/y for the period 2002 to 2005 (Chen et al., 2006); and -84 ± 5 Gt/y for the period 2003 to 2008 (Luthcke et al., 2008).

7.3.2.3. Canadian Arctic

The Canadian Arctic region includes the Queen Elizabeth Islands (including Ellesmere, Devon, and Axel Heiberg Islands), and Bylot and Baffin Islands. Glaciers cover an area of about 150 000 km^2 in this region. Current and past glacier extents are known from aerial photography taken during the period 1956 to 1961 and from satellite imagery collected since the mid-1970s (and especially since 1999) (Sharp et al., 2003, in press; Dowdeswell et al., 2007; Wolken et al., 2008). Comparison of ice extents from around 1960 with trimline and end moraine

positions visible on satellite imagery shows that there has been extensive ice retreat across the region from glacier maximum positions reached by the end of the Little Ice Age. The ice-covered area on the Queen Elizabeth Islands decreased by 37% from a Little Ice Age maximum of 170 124 km^2 to about 107 735 km^2 in 1960, and all permanent ice was removed from most of the western islands (Dyke, 1978; Wolken et al., 2008). Further glacier retreat has occurred since 1960, with ice area reductions by 2000/2001 of 5.1% (253 km^2) on Bylot Island (Dowdeswell et al., 2007) and 2.7% (2844 km^2) on the Queen Elizabeth Islands (Sharp et al., 2003). The 14 400 km^2 Devon Island Ice Cap decreased in area by 338 ± 40 km^2 (2.4%) (Burgess and Sharp, 2004). The area of small plateau ice caps in interior northern Baffin Island decreased by 55% (from 150 km^2) between 1958 and 2005 (Anderson et al., 2008), while the area of 264 glaciers on the Cumberland Peninsula in southeast Baffin Island decreased by about 294 km^2 (12.5%) between around 1920 and 2000 (Paul and Svoboda, 2009). In the latter study, proportional area reductions were greatest for the smallest glaciers, and the rate of area loss roughly doubled after 1975.

Glaciological mass balance measurements have been made in the Queen Elizabeth Islands since the late 1950s (Cogley et al., 1996; Koerner, 2005) (Figure 7.10), but there are no long-term records from Bylot or Baffin Islands, and no regular measurements since the mid-1980s (Hooke et al., 1987). In this region, variability in the annual mass balance arises almost entirely from variability in the summer balance (Koerner, 2005). The cumulative surface mass balance of all measured glaciers in the region has been negative over the measurement period (e.g., Devon Island Ice Cap -4.4 m w.e. from 1961 to 2007, White Glacier, Axel Heiberg Island -6.8 m w.e. from 1960 to 2007). Compared with the 1963 to 1987 period, the rate of mass loss accelerated significantly after 1987 (Gardner and Sharp, 2007). There have been a number of very negative mass balance years since 1997 (notably 1998, 2001, 2005, 2007, 2008, 2009) that suggest a further acceleration of mass loss since 1997. The cumulative mass loss from the Devon Island Ice Cap from 1997 to 2007 exceeds that for the period 1963 to 1997. Consistent with this trend, geodetic measurements on the Barnes Ice Cap suggest a progressive increase in rates of thinning from 0.12 m/y (1970 to 1984), to 0.76 ± 0.35 m/y (1984 to 2006), and 1.0 ± 0.14 m/y (2004 to 2006) (Sneed et al., 2008).

Iceberg calving also contributes to mass loss from many of the larger ice caps (Burgess et al., 2005). Separate estimates of surface mass balance and iceberg calving fluxes are available for two large ice caps in Arctic Canada: the Devon Island Ice Cap and Prince of Wales Icefield. For the Devon Island Ice Cap, the estimates equate to an overall mean annual balance of either -2.3 Gt/y (1963 to 2000) or -5.2 ± 1.7 Gt/y (1980 to 2006), depending on whether the net surface balance estimates of Mair et al. (2005) (-1.8 ± 0.8 Gt/y) or Gardner and Sharp (2009) (-4.7 ± 1.7 Gt/y) are added to the Burgess et al. (2005) estimate of mass loss due to iceberg calving (-0.5 Gt/y). The surface mass balance of the Prince of Wales Icefield (1963 to 2003) appears to be within an error of zero, so the overall negative balance (-2.0 ± 0.5 Gt/y) is similar to the annual calving flux (1.9 ± 0.2 Gt/y) (Mair et al., 2009).

Repeat airborne laser altimetry flown over some of the larger ice caps in the region in 1995 and 2000 showed that most ice caps were thickening or maintaining a constant thickness at elevations above 1600 m, and thinning at lower elevations (Abdalati et al., 2004). Rates of thinning were greater on Baffin Island ice caps (>1 m/y) than in the Queen Elizabeth Islands (< 0.5 m/y). Based on these data, Abdalati et al. (2004) suggested that the total mass balance of Arctic Canada in the period 1995 to 2000 was -22.9 Gt/y.

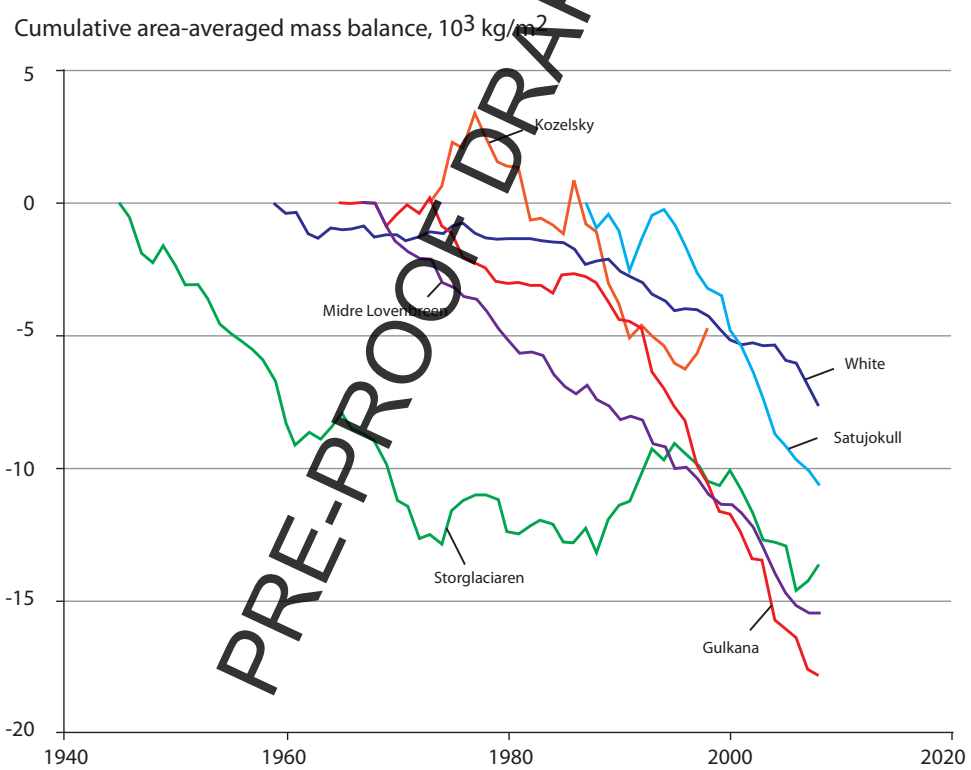


Figure 7.10. Cumulative area-averaged mass balance for six Arctic glaciers: White Glacier (Axel Heiberg Island), Gulkana Glacier (Alaska), Storglaciären (Sweden), Midre Lovénbreen (Svalbard), Satujökull (Hofsjökull, Iceland), and Kozelsky (Kamchatka) over the period of record. All records show a negative cumulative balance (net thinning) over the period of record and more rapid thinning after about 1990 (although the precise timing of this varies between regions). Source: AMAP.

Thus, measurements of glacier extent, thickness, and surface mass balance are all consistent in showing progressive and possibly accelerating shrinkage of the Canadian Arctic land ice cover. Nonetheless, a precise and regionally comprehensive estimate of the magnitude of these changes and their implications for global sea level is not yet available.

7.3.2.4. Greenland (mountain glaciers and ice caps)

Glaciers occupy about one quarter of the peripheral area around the Greenland Ice Sheet (Thomsen and Weidick, 1992) (Figure 7.11). The glaciers of Disko Island (Qeqertarsuaq) and the Nuussuaq and Svartenhuk peninsulas, West Greenland, that covered an area of 3832 km^2 in 2001, lost between 16% and 21% of their area from the end of the Little Ice Age to 2001 (Citterio et al., 2009). There were no noticeable changes in the position of most calving glacier fronts between the late 19th century and the mid-1980s (Weidick, 1995). Between 1978 and 1991, major tidewater glacier termini along the northern coast of Kangerlussuaq and the southern part of Blosseville Kyst showed little change or a slight loss (0.1 to 0.5 km^2) in the glacier tongue area (Dwyer, 1995).

Extreme fluctuations of the terminus of the surge-type tidewater glacier Sortebrae involved a retreat of about 1.5 km between 1933 and 1943, a surge advance of about 10 km by 1950, a retreat of 8 km between 1950 and 1981, followed by stagnation between 1981 and 1992, and a second surge advance of over 5 km between 1992 and 1994 (Jiskoot et al., 2001). Another example of surge-related terminus fluctuations is shown by a small land-based Scoresby Sund glacier, which retreated 0.5 km between 1987 and 2000, and advanced by 2.8 km between 2000 and 2007 (Jiskoot and Juhlin, 2009). Neighboring glaciers did not change their terminus positions significantly during the same period.

7.3.2.5. Iceland

Glacier terminus variations in Iceland have been monitored regularly since 1930. Most non-surge-type glaciers retreated rapidly from 1930 to 1960, but more slowly in the 1960s. Glaciers began to advance in the 1970s and rates of advance peaked in the mid-1980s. In the 1990s, glaciers began to retreat again and all monitored glaciers were retreating by 2000. Terminus fluctuations follow changes in mean summer temperature with a delay of only a few years (Sigurðsson et al., 2007).

Mass balance measurements were initiated on Hofsjökull in 1987/1988 (Figure 7.10; Satujökull), Vatnajökull in 1991, and Langjökull in 1996. Vatnajökull's balance was positive from 1991 to 1994 and near zero in 1994 to 1995, but it has been negative ever since, with a cumulative specific balance of about -9.2 m w.e. (84 Gt, or 2.7% of the ice cap mass) from 1994 to 2006 (Björnsson and Pálsson, 2008). In the 1990s, mass loss from Vatnajökull due to basal melting resulting from geothermal activity and volcanic eruptions averaged 0.55 Gt/y, about 4% of the total surface ablation in a year of zero surface balance. However, the Gjalp volcanic eruption in 1996 melted about 4.0 Gt of ice. The net balance of Langjökull has been negative since measurements began and the cumulative specific net balance from 1996 to 2006 was about -12.8 m w.e. (13.1 Gt, 7% of the ice cap mass) (Björnsson and Pálsson, 2008). The current period of negative mass balance reflects a combination of high summer temperatures and low winter snowfall.

Figure 7.11. Mountain glaciers and ice caps occupy much of the peripheral areas of Greenland that lie beyond the margins of the Greenland Ice Sheet. Source: Anthony Arendt, University of Alaska.

7.3.2.6. Scandinavia

After their Little Ice Age maximum, Scandinavian glaciers retreated slightly until the 1930s/1940s, when there was a period of more rapid retreat that lasted into the 1950s for maritime outlet glaciers and the 1970s for more slowly responding outlets in the interior. Increased winter precipitation between 1988/1989 and 1994/1995 triggered a readvance of some glaciers in western Norway and Sweden, but higher summer air temperatures after 2001 induced an episode of rapid glacier retreat (Andreassen et al., 2005; Nesje, 2005, 2009). The longest mass balance records from Scandinavia date to the late 1940s (e.g., Holmlund et al., 2005), but the number of measurements increased in the early 1960s. Since that time, glaciers in inland Scandinavia (for which net balance variability is due largely to variability in the summer balance) have had a cumulative negative mass balance (Figure 7.10), while maritime glaciers in western Norway and some others such as Storglaciären in Sweden (for which net balance variability is due largely to variability in the winter balance) have had a cumulative positive net balance (especially in the 1990s when winter precipitation was unusually high) (Nesje, 2009). Since 2001, however, the mass balance of Norwegian glaciers has become predominantly negative. The North Atlantic Oscillation has a strong influence on the variability of the mass balance of maritime glaciers in western Norway (Reichert et al., 2001).

7.3.2.7. Svalbard

Over the past century, there has been a general trend toward retreat and thinning of Svalbard glacier fronts (Dowdeswell, 1986), excluding those that have surged during the period of observation, and thickening of higher elevation regions (Bamber et al., 2004; Nuth et al., 2007, 2010; Kohler et al., 2007; Moholdt et al., 2010a,b). Mass balance has been measured on two Svalbard glaciers (Austre Bröggerbreen and Midre Lovénbreen) since 1967 and on two more (Kongsvegen and Hansbreen) since the late 1980s (Hagen et al., 2003b). Almost all annual net balances are negative, the most notable exceptions

being in 1987 and 1991 when all the measured balances were positive. There is no significant long-term trend in any of the annual balance records. Large annual variations are driven mainly by variations in the summer ablation. The cumulative balances of Austre Bröggerbreen (1967 to 2007) and Midre Lovénbreen (1968 to 2007) were -20.3 and -15.3 m w.e., respectively, while that of Hansbreen (1989 to 2007) was -6.2 m w.e. (WGMS, 2009). Kongsvegen had a much smaller negative annual balance and a cumulative balance of -1.6 m w.e. (1987 to 2007) (WGMS, 2009), indicating that larger glaciers with high-elevation accumulation areas are experiencing more favorable mass balance conditions than smaller, low-lying glaciers. This is also the case for Austfonna in northeastern Svalbard, which had a surface net balance of -0.05 m/y w.e. for the five-year period 2004 to 2008 (Moholdt et al., 2010a). Shallow ice cores from Austfonna also indicated a net surface mass balance close to zero for the period between the Chernobyl fallout in 1986 and 1999 (Pinglot et al., 1999).

Hagen et al. (2003b) used the hypsometry (area-altitude distribution) of the glaciers and curves of net balance versus altitude for 13 regions to estimate the surface balance of Svalbard glaciers over the period 1968 to 1998. They found it to be slightly negative (-0.5 ± 0.1 Gt/y), or a mean specific surface net balance of -0.014 ± 0.003 m w.e./y. Mass loss by iceberg calving was estimated to be -4.0 ± 1.0 Gt/y (Dowdeswell and Hagen, 2004), making the total net balance of the archipelago significantly more negative, in total -4.5 Gt/y, or -0.12 ± 0.03 m w.e./y. A more recent estimate of the Svalbard calving flux (2000 to 2006) is -6.8 ± 1.7 Gt/y (Błaszczyk et al., 2009).

Geodetic measurements suggest that the rate of volume loss from a number of glaciers, especially low-lying glaciers close to the west coast, has increased by up to a factor of four since the early 1960s (Kohler et al., 2007). This is consistent with summer temperatures recorded at Longyearbyen, which increased by 0.04 °C/y from 1967 to 2006 and 0.17 °C/y from 1997 to 2006 (Førland and Hanssen-Bauer, 2003). Nuth et al. (2010) compared satellite altimetry from ICESat (2003 to 2007) to older topographic maps and digital elevation models (1965 to 1990) and estimated the geodetic mass change over the past 40 years for Svalbard, excluding Austfonna and Kvitøya, to be -9.7 ± 0.6 Gt/y (surface mass balance of -0.36 ± 0.02 m w.e./y). When the Austfonna estimate is added, this gives an overall mass balance for Svalbard of nearly -13 Gt/y. Moholdt et al. (2010b) used repeat track ICESat laser altimetry to assess mass and volume changes of Svalbard glaciers from 2003 to 2008. During this period, most regions experienced thinning at low elevations and thickening at high elevations. The geodetic mass balance for all of Svalbard's glaciers (not taking into account the effects of advance and retreat of calving glacier termini) was estimated to be -4.3 ± 1.4 Gt/y. The largest mass losses were in western and southern Svalbard, while Austfonna and glaciers in northeastern Spitsbergen gained mass over the period. Thus, available estimates of Svalbard net mass balance range from -4.3 to -14 Gt/y, depending upon the period of measurement and which calving flux estimate is used.

7.3.2.8. Russian Arctic islands

The major glacierized regions in the Russian Arctic are in the archipelagos of Franz Josef Land, Severnaya Zemlya, and Novaya Zemlya. Between 1952 and 2001, the glacierized areas decreased by 2.7% on Franz Josef Land, 0.4% on Severnaya Zemlya, and 1.4% on Novaya Zemlya. The total area reduction was about 725 km^2 (-375 km^2 on Franz Josef Land, -284 km^2 on Novaya Zemlya, and -65 km^2 on Severnaya Zemlya) (Glazovsky and Macheret, 2006). In Novaya Zemlya, net glacier recession has been greater on the west coast than on the east coast (Glazovsky, 2003). Using local area-volume scaling relationships, volume losses from the three archipelagos between 1952 and 2001 were estimated at -71 km^3 (Franz Josef Land), -24 km^3 (Severnaya Zemlya), and -150 km^3 (Novaya Zemlya). Assuming the volume was lost as ice, this gives a mean annual mass loss of -4.5 Gt/y from the three archipelagos (Glazovsky and Macheret, 2006).

Limited measurements of the surface mass balance of Shokal'sky Glacier on Novaya Zemlya, supplemented by reconstructions based on regression analyses between winter temperature / accumulation and summer temperature / ablation, suggest predominantly negative balances between 1930 and the mid-1990s (Zeeberg and Forman, 2001). Direct measurements on Vavilov Ice Cap (Severnaya Zemlya) between 1974 and 1988 also showed a slightly negative surface balance (-0.03 m w.e./y) (Barkov et al., 1992).

7.3.2.9. Russian mountains

The USSR glacier inventory mapped the glaciers of northeastern Siberia from aerial photography. Glacier changes have been determined by comparison with 2002/2003 Landsat7 ETM+ imagery. The area of glaciers of the Suntar Khayata mountain range (Figure 7.12) decreased from 200 km^2 in 1945 to 162 km^2 in 2002/2003, while that of glaciers in the Chersky mountain range decreased from 156 km^2 in 1970 to 113 km^2 in 2002/2003 (Ananicheva et al., 2006). The area of the Byrranga glaciers on the Taymir Peninsula decreased from 30.5 km^2 in 1967 to 24.4 km^2 in 2003, a reduction of 20%. South- and east-facing glaciers had the largest areal reductions, consistent with greater warming trends and smaller influence of westerly-derived cyclones on the eastern side of the mountain range (Ananicheva and Kapustin, in press a).

On the Koryak Upland, a peninsula between Kamchatka and Chukotka, 248 glaciers were mapped from maps and aerial photographs dating from the 1950s and using Landsat 7 ETM+

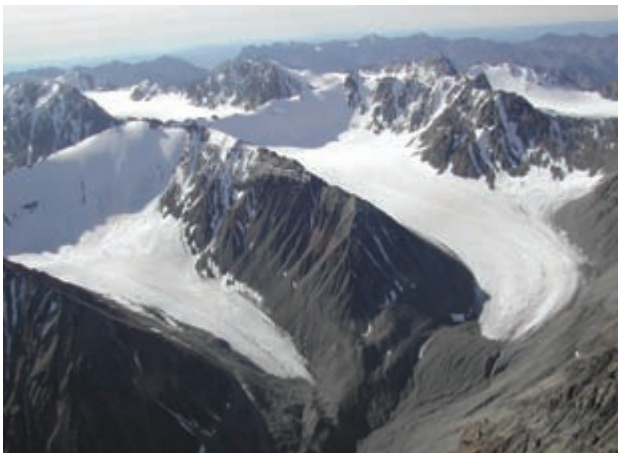


Figure 7.12. Mountain glaciers in the Suntar Khayata Range of northeastern Siberia. Note the well-developed meltwater drainage systems on the glacier surfaces. Source: Maria Ananicheva, Russian Academy of Science.

and ASTER data from 2003. Their combined area decreased by 66.5%, from 176.6 km^2 to 54.4 km^2, over the intervening period. Glaciers with a northeasterly aspect showed the largest proportional area reductions. These glacier area changes are the largest recorded in the Russian sub-Arctic and reflect a response to a combination of climate warming and decreased snowfall from Pacific sources (Ananicheva and Kapustin, in press b).

At the end of the 1950s, there were 143 glaciers in the Ural Mountains with a total area of 28.7 km^2 (Osipova, 2006). From 1958 to 2001, mass balance observations were carried out on two Polar Ural glaciers (IGAN Glacier, area 0.73 km^2, and Obruchev Glacier, area 0.35 km^2). These records have been extended to cover the entire 20th century by correlation with local measurements of summer air temperature and winter precipitation. The resulting series show periods of negative balance between 1900 and 1920, in the late 1950s and 1960s, and since 1980. Positive balances occurred between 1920 and the mid-1950s (Kononov et al., 2005). From 1953 to 2000 (1981 for IGAN Glacier), the average area decrease rate of the four glaciers in the region (Obruchev, MGU, Chernov, IGAN) was 0.9% per year (Glazovsky et al., 2007).

7.3.2.10. Kamchatka

There are currently 448 glaciers in Kamchatka, with an area of about 874 km^2. The climate of Kamchatka is strongly influenced by intense storm activity from the Pacific Ocean, the thermal influence of the Sea of Okhotsk on the western side of the Peninsula, and the topography of the region. The mass balance of Kozelsky Glacier was positive from 1973 to 1978 and generally negative thereafter, with a cumulative mass balance of about -5 m w.e. over the period of record (1973 to 1997; Dyurgerov and Meier, 2005) (Figure 7.10).

7.3.2.11. Summary of regional changes

In general, glaciers across the Arctic began retreating and losing mass in the early 20th century. Area changes and mass loss continued through the mid-20th century, with some regions experiencing brief episodes of slower retreat, reduced mass loss, or even mass gain. The rate of mass loss increased during the past decade across most regions of the Arctic. This marks a change from the situation reported by Dowdeswell et al. (1997), who found that although most Arctic glaciers had experienced predominantly negative net surface mass balance over the previous few decades, there was no uniform trend in mass balance.

7.3.3. Links between mass balance and climate

Annual glacier mass balances often correlate over spatial scales as large as 1200 km (Cogley and Adams, 1998; Rasmussen, 2004) as a result of trends in climate that are consistent over broad regions. These trends often occur due to synoptic-scale teleconnections between the positioning of the polar jet stream, the phase of the El Niño Southern Oscillation (ENSO), and regional patterns in temperature and precipitation. The mass balance of glaciers in Alaska, western Canada, and the U.S. Pacific Northwest is strongly affected by a decadal ENSO-like phenomenon, otherwise known as the Pacific Decadal Oscillation (PDO), which is measured by variations in the sea-surface temperature of the Gulf of Alaska (Bitz and Battisti, 1999; Josberger et al., 2007). Mass balance measurements from Gulkana Glacier in interior Alaska, Wolverine Glacier in maritime Alaska, and South Cascade Glacier in coastal Washington State show the complex reaction of glacier mass balance to variations in the PDO that result in major changes in the positioning of the North Pacific storm track and the distribution of winter precipitation. From 1966 to 1977, the PDO was in a strong cold phase and the net balance of the three glaciers was slightly negative. In 1977, the PDO switched into a strong warm phase, and the mass balance of South Cascade became very negative, a trend that continues to present. The mass balance of Gulkana and Wolverine Glaciers began to increase as the storm track was diverted toward Alaska. From 1987 to present, the PDO warm phase has weakened, and the net balance of all three glaciers has become uniformly negative.

Gardner and Sharp (2007) showed that accelerated glacier surface mass loss in the Canadian Arctic after 1987 was associated with increased July mean air temperature and a shift in the mean position of the center of the July circumpolar vortex from the western hemisphere to the eastern hemisphere. This allowed increased intrusion of warm air from continental North America into the Queen Elizabeth Islands in summer. July vortex types associated with warm temperatures and negative mass balance anomalies were also more common from 1948 to 1962, so predominantly negative surface mass balances may also have characterized the 1950s.

Occasional periods of positive balance on Novaya Zemlya (especially in the late 1940s and early 1950s) occurred when the North Atlantic Oscillation (NAO) was in a positive phase, sea-surface temperatures in the southern Barents Sea were anomalously warm, and winter precipitation was increased (Zeeberg and Forman, 2001). Negative surface balances occurred whenever summer temperatures rose more than 1 °C above the 1933 to 1991 mean, most notably from the early 1950s to the late 1960s. Positive NAO conditions also coincided with the period of positive mass balances in western Scandinavia in the early 1990s (Pohjola and Rogers, 1997).

7.4. Proxy indicators of surface mass balance

- The number of *in situ* measurements of the surface mass balance of glaciers in the Arctic is small, and the distribution of monitored glaciers is uneven. Therefore, it is difficult to estimate accurately the surface mass balance of the regional ice cover (the quantity that needs to be known to determine the contribution of Arctic glaciers to global sea level change) on the basis of these measurements alone.
- There is, therefore, considerable interest in using remote sensing methods to repeatedly measure regional-scale patterns in parameters that vary in ways that are well-correlated with surface mass balance in order to generate records that may reflect the temporal trends and variability in surface mass balance.
- Parameters of interest include the length of the summer melt season, the distribution of snow and ice facies across mountain glaciers and ice caps, the equilibrium-line

altitude, the accumulation area ratio, the surface albedo, and the surface temperature. These can be measured using active microwave methods or passive monitoring in the visible and infrared parts of the electromagnetic spectrum. To date, such measurements have been made more widely on the large ice sheets than on mountain glaciers and ice caps, but this is changing with improvements in sensor resolution.

- Pan-Arctic mapping of summer melt season duration and the end-of-summer distribution of snow and ice facies has been possible for the 1999 to 2009 decade due to the availability of enhanced resolution data products from the Seawinds scatterometer on QuikSCAT. That satellite is no longer functional and there is as yet no new data series to replace the one it generated.

7.4.1. Introduction

Glaciological (*in situ*) measurements of the surface mass balance of Arctic glaciers and ice caps are limited to a small number of glaciers that are unevenly distributed across the Arctic and are biased toward small, land-terminating glaciers. For many of these glaciers, the time series of surface mass balance measurements are discontinuous, of short duration, and not recent. In some regions of the Arctic, *in situ* surface mass balance data are simply nonexistent. Such limitations are mainly due to the remoteness of the Arctic and the intense logistical demands associated with establishing and maintaining *in situ* surface mass balance observations. However, common practice has been to produce regional-scale estimates of surface mass balance based on interpolation of the small number of *in situ* observations. This results in large uncertainties where data are sparse, so that it is not clear whether these estimates truly represent regional surface mass balance trends over the Arctic. Because of these limitations, there is a pressing need to develop alternative methods for estimating trends in surface mass balance and to be able to apply them repeatedly over large areas in order to monitor changes in surface mass balance at regional scales.

Proxy indicators of surface mass balance are parameters or indices that are directly related to the surface mass balance process and that can be used to infer changes in surface mass balance. For example, variations in the abundance of melt layers in ice cores have been used for this purpose (e.g., Koerner and Fisher, 1990; Kinnard et al., 2008). Currently, regional-scale proxy indicators of annual surface mass balance are typically developed from satellite-derived remote sensing data. Some of these proxies provide an indication of the sign of the surface mass balance change between consecutive mass balance years (e.g., Wolken et al., 2009), while others may be useful in the development of methodologies for up-scaling *in situ* surface mass balance measurements to the regional-scale glacier cover (e.g., Wang et al., 2005) or for validating predictions from surface mass balance models.

7.4.2. Proxy indicators, data sources, and methods of measurement

7.4.2.1. Melt duration

Spatial and temporal variability in the surface mass balance of many Arctic glaciers is controlled by the variability in surface melt (Koerner, 2005; Gardner and Sharp, 2007). Surface melt characteristics (extent and duration) can provide valuable information about the melt process and can be detected at regional scales by satellite remote sensing. Because melt usually occurs in all regions of Arctic glaciers and ice caps every summer, melt extent provides little information about variability in the summer melt process on these ice masses. Melt duration is a much better descriptor and is well correlated with the annual positive degree-day total (a parameter used to simulate snow and ice melt rates in temperature index melt and mass balance models) (Wang et al., 2005), and with summer mean temperatures and geopotential heights in the lower troposphere as derived from climate re-analyses (Wang et al., 2005, 2007; Sharp and Wang, 2009).

Microwave remote sensing is well suited for mapping melt duration on Arctic ice masses. Passive microwave remote sensing is commonly used for melt mapping on Greenland, but its coarse spatial resolution (≥25 km) precludes its use on the smaller Arctic glaciers and ice caps. Data from Ku- and C-band microwave scatterometers have been used successfully to map annual melt duration on ice caps and glaciers in the Arctic (Smith et al., 2003; Wang et al., 2005; Sharp and Wang, 2009; Sharp and Wolken, 2009, 2010; Wolken et al., 2009) (Figure 7.13). Annual melt duration is determined by calculating the number of days between the melt onset and freeze-up dates minus the duration of any intervening periods when melt was not detected. Wang et al. (2005) used time series of enhanced resolution data from the SeaWinds scatterometer aboard QuikSCAT (Early and Long, 2001; Long and Hicks, 2005) to determine melt onset and freeze-up dates and to determine annual melt season durations on ice caps in Arctic Canada (2000 to 2004). On these ice caps, annual mean melt duration varies with ice surface elevation (r = -0.80) and distance from Baffin Bay (r = -0.44) and is strongly linked to variations in geopotential height at the 500 hPa level. For the period 2000 to 2004, average melt duration for individual ice caps ranged from 26 days on the Agassiz Ice Cap to 62 days on the Manson Icefield. For the Queen Elizabeth Islands as a whole, the shortest melt season was 2002 and the longest was 2001.

In the Eurasian Arctic (Svalbard, Severnaya Zemlya, Novaya Zemlya), average melt duration (2000 to 2004) is well correlated with latitude, longitude, and elevation and is consistent with the steep climatic gradient across the region (i.e., warm and moist in the west to cool and dry with persistent sea ice in the east; Dowdeswell et al., 2002; Sharp and Wang, 2009). Average melt duration on Severnaya Zemlya (51 days) is significantly shorter than on Novaya Zemlya (75 days) and Svalbard (77 days). Data show that 2000 was the shortest melt season and 2001 the longest during the 2000 to 2004 period (Sharp and Wang, 2009). Variations in annual mean melt duration were strongly correlated with the mean (June and August) 850 hPa air temperature.

7.4.2.2. Snow and ice facies

Glacier surfaces are organized into zones on the basis of differences in the physical properties of surface and near-surface snow and ice (Benson, 1962). The differences between snow and ice facies are related to variations in the magnitude

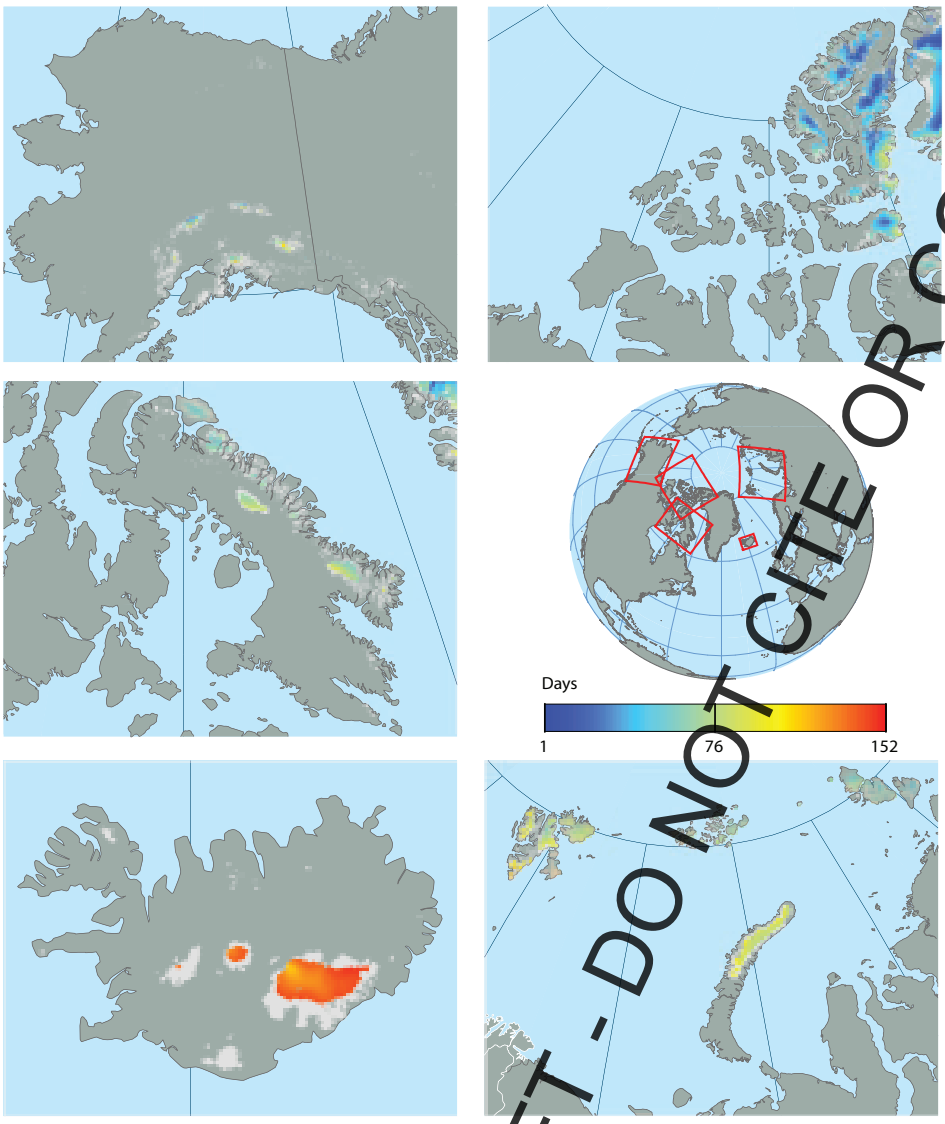


Figure 7.13. Mean annual melt duration (days) on larger Arctic ice masses for the period 1999 to 2008 based on QuikSCAT scatterometer data. Source: Gabriel Wolken, Alaska Division of Geological and Geophysical Surveys.

of summer melt, which are strongly correlated with elevation. Snow and ice facies on glaciers and ice caps in the Arctic include the following (from high to low elevation): dry snow facies – formed where no melt occurs and snow is converted to ice solely by compaction; percolation facies – formed where meltwater percolates into the snowpack and refreezes as lenses, layers and pipes, but the snowpack never reaches the melting point throughout the melt season; saturation facies – formed where the temperature of the entire snowpack reaches the melting point sometime during the melt season and wet snow refreezes at the end of the melt season; superimposed ice facies – formed in a transitional zone in the lowest part of the accumulation area where meltwater freezes directly onto the glacier ice surface; and glacier ice facies – exposed in the ablation zone in summer when summer melt removes the annual snow accumulation and exposes glacier ice at the surface (Figure 7.14). There is often a transitional zone between snow and ice facies in which a mixture of these facies can be found. The distribution of snow and ice facies can change annually, and not all facies are found on all ice masses in all years. Because the distribution of snow and ice facies is directly linked to the balance between the processes of accumulation and surface melt, it can provide valuable proxy information about spatial and temporal variations in the surface mass balance.

Snow and ice facies have been mapped at regional scales using optical and microwave remote sensing. Mapping with optical sensors has had limited success, largely due to the often subtle contrast between snow and ice facies and to the occurrence of fresh snow (which conceals the underlying facies) and cloud cover (which conceals the glacier) at the end of the ablation season (Østrem, 1975a; Nolin and Payne, 2007; Dowdeswell et al., 2007). Microwave radiation is unaffected by cloud cover, but is sensitive to the surface and volumetric physical properties

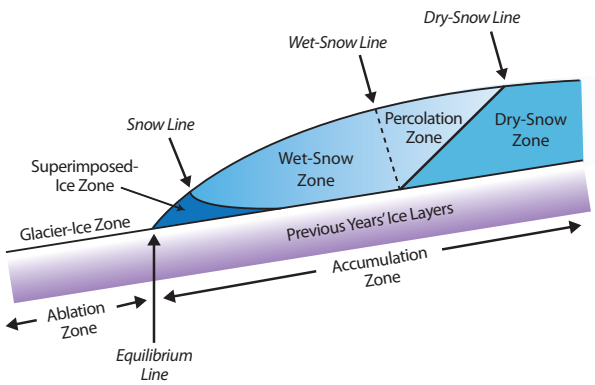


Figure 7.14. Snow and ice facies zones on a glacier or ice cap. Source: after Benson (1962).

of snow and ice. Snow and ice facies on Arctic glaciers and ice caps have been linked to specific signatures acquired by active microwave sensors, and mapped on the glaciers Kongsvegen and Lovénbreen (Svalbard) using synthetic aperture radar (SAR; ERS-1/2) data (Engeset et al., 2002; König et al., 2002). Firn, superimposed ice, and glacier ice facies were identified in these studies, but the complicated nature of backscatter in the superimposed ice zone precluded the identification of the equilibrium line. Further work on Kongsvegen using *in situ* data and SAR concluded that the superimposed ice zone represents 35% of the glacier (Obleitner and Lehning, 2004). Langley et al. (2008) and Brandt et al. (2008) also investigated snow and ice facies on Kongsvegen by combining data from SAR and ground-penetrating radar (GPR) and concluded that SAR zones were in very good agreement with GPR-derived glacier facies. On Austfonna (Svalbard), Dunse et al. (2009) used GPR to determine the multi-year (2004 to 2007) snow and ice facies distribution and linked variability in facies over the study period to changes in surface mass balance.

Regional-scale mapping and monitoring of snow and ice facies have been performed using microwave scatterometer data. Wolken et al. (2009) used a classification of end-of-summer, enhanced resolution (2.225 km^2) data from QuikSCAT (tuned using available field observations) to map the distribution of snow and ice facies in the Canadian High Arctic and its interannual variability (1999 to 2005) (Figure 7.15). Increases in summer air temperature and melt duration were associated with a decrease in the areas of the percolation and saturation zones, an increase in the area of the glacier ice zone, and an increase in the elevation of inter-facies boundaries.

7.4.2.3. Equilibrium-line altitude

The equilibrium line is the line connecting points on a glacier surface where annual accumulation equals annual ablation. Above the equilibrium line (the accumulation area) there is a net gain in mass for a particular year, whereas below the equilibrium line (the ablation area) there is a net loss in mass. The equilibrium-line altitude (ELA) is the average altitude of the equilibrium line at the end of the ablation season in any balance year. The ELA is related to the surface mass balance because it rises and falls as the annual mass balance of the whole glacier becomes more or less negative (Østrem, 1975a; Hagen and Liestøl, 1990). It can be used to identify variability and trends in surface mass balance when monitored annually. The long-term ELA represents the average position of the ELA over many years. Change in the long-term ELA, generally linked to a change in glacier geometry, serves as an important indicator of regional climate change.

The ELA is most accurately determined using standard *in situ* glaciological techniques for measuring surface mass balance. Dyurgerov (2002), Dyugerov and Meier (2005) and Dyurgerov et al. (2009) compiled all available time series (around 280) of *in situ* mass balance data and related variables, including ELA, for glaciers outside the two ice sheets. For the few glaciers in the Arctic where ELAs are determined through *in situ* measurements, ELA time series are characterized by a high degree of interannual variability. However, while these measurements are robust and invaluable to the glaciological community, they are also spatially restricted and limited in their ability to provide a meaningful measure of ELA variability on

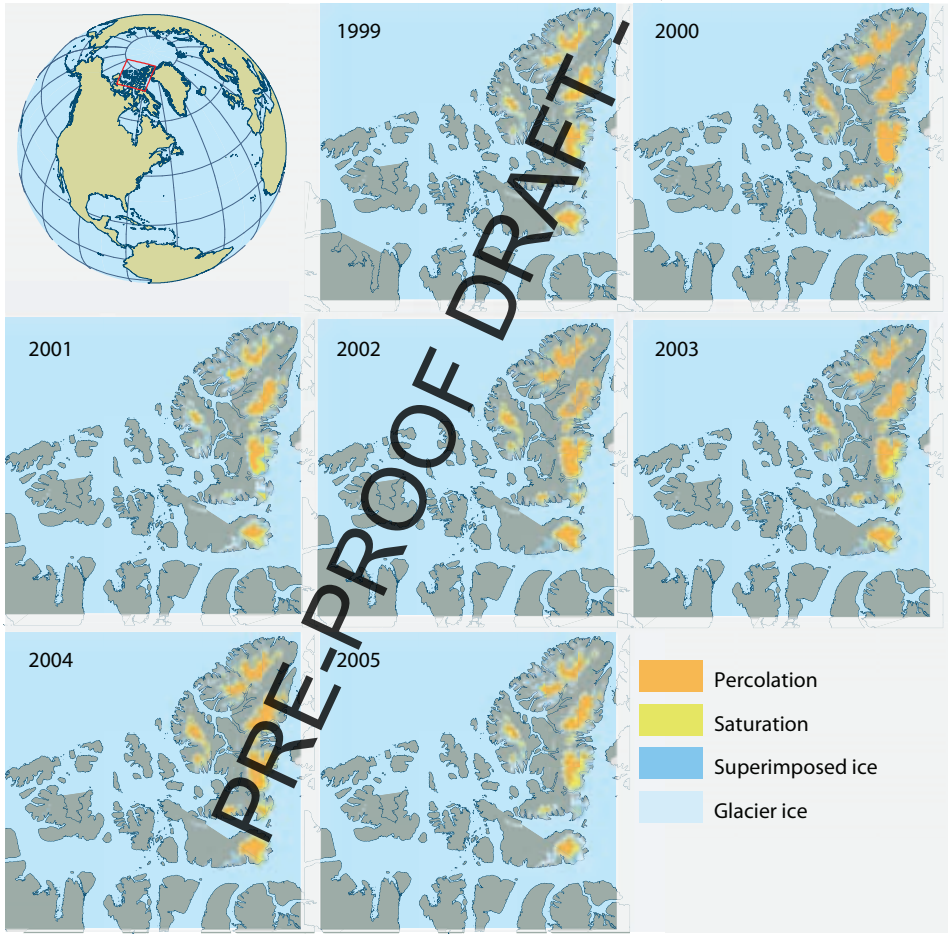


Figure 7.15. Annual distribution of snow and ice facies at the end of the melt season on ice caps in the Canadian High Arctic (1999-2005) as derived from QuikSCAT scatterometer data. Taken from Wolken et al. (2009). Permission from AGU is required.

a regional scale.

Alternative methods have been developed for monitoring regional-scale changes in the annual ELA, the most effective of which involve the use of satellite microwave remote sensing. In Arctic Canada, regional-scale variability in the ELA can be inferred from changes in the lower boundary of the superimposed ice zone, which is highly correlated with changes in surface mass balance measurements (Wolken et al., 2009). Scatterometer-derived changes in the annual average ELA for each ice cap show a generally coherent pattern of change during the study period (1999 to 2005), with the highest ELAs occurring in 2001, the year with the longest melt season.

7.4.2.4. Accumulation area ratio

The accumulation area ratio (AAR) is the ratio of the area of the accumulation area to the area of the entire glacier. Because the lower limit of the accumulation area is determined by the position of the equilibrium line, the AAR and the ELA are directly related. If a record of the AAR is combined with *in situ* surface mass balance data, then it is possible to produce quantitative estimates of changes in the surface mass balance with respect to changes in the AAR (Dyurgerov et al., 1996, 2009; Hock et al., 2007b). Quantification of regional-scale changes in surface mass balance from AAR changes in the Arctic is limited by the sparseness of *in situ* surface mass balance data for use in calibration. However, regional-scale changes in surface mass balance can be qualitatively assessed based on the relative changes in the AAR between regions. Annual AARs for ice caps in the Canadian High Arctic were determined from maps of the distribution of snow and ice facies derived from QuikSCAT scatterometer data for the period 1999 to 2005 (Wolken et al., 2009). Results from this study indicate considerable spatial and temporal variability in the annual AAR during this period. For the Queen Elizabeth Islands as a whole, the mean AAR was 0.75 over the period 1999 to 2005. The smallest mean AAR (0.50) occurred in 2001and the largest (0.82) in 2004 (Wolken et al., 2009).

7.4.2.5. Albedo

The albedo (the fraction of incoming solar radiation reflected by a surface) of snow and ice is a function of snow/ice grain size, surface water content, solar incidence angle (Wiscombe and Warren, 1980), and snow impurities (Warren and Wiscombe, 1980), including black carbon from natural and industrial sources (McConnell et al., 2007; Forsström et al., 2009). On Arctic glaciers, solar radiation is usually the main source of energy for surface melt, and its magnitude and distribution are controlled in part by the glacier surface albedo. As a result, and because factors such as grain size and water content are coupled to surface melt rates, interannual variations in surface albedo and their spatial patterns over individual glaciers can serve as a proxy indicator of surface mass balance.

In situ observations of albedo can be scaled-up with satellite optical remote sensing measurements (Williams, 1987; Reijmer et al., 1999; Greuell and Knap, 2000; Nolin and Payne, 2007), producing data that can be used to identify reductions in albedo that provide an indication of melt onset (Hall et al., 1987; Bindschadler et al., 2001). De Ruyter de Wildt et al. (2002, 2003) used Advanced Very High Resolution Radiometer (AVHRR) data for Vatnajökull, Iceland to determine the evolution of the surface albedo over the melt season and estimate the potential absorbed radiation, which was shown to correlate well with the mean specific mass balance ($r = 0.87$ and 0.94 for different outlet glaciers). This methodology was further developed by Greuell et al. (2007), who generated estimates of mass balance for the whole of Vatnajökull.

7.4.2.6. Land (glacier) surface temperature

Because winter precipitation is consistently low in the Arctic (excluding southern Alaska, Scandinavia, Iceland, and Kamchatka), variability in glacier mass balance is largely controlled by summer temperature. While precipitation events during summer can be important as summer snowfalls result in short-term increases in surface albedo, temperature tends to drive the ablation process and control variations in surface mass balance (Koerner, 2005; Gardner and Sharp, 2007). In the absence of systematic *in situ* temperature observations, which are available for only a few areas across the Arctic, satellite-borne thermal infrared sensors can be used to monitor clear-sky land surface temperature (LST) and surface melt at regional scales over the terrestrial cryosphere (Key and Haefliger, 1992; Key et al., 1997; Stroeve and Steffen, 1998; Hall et al., 2004, 2006, 2008a,b). In the Eurasian Arctic, Sharp and Wang (2009) used Moderate-Resolution Imaging Spectroradiometer (MODIS) LST data in place of *in situ* near-surface air temperature data to tune melt detection algorithms applied to QuikSCAT scatterometer data. However, there has been little other use of LST data on Arctic glaciers and ice caps. This is probably because reliable satellite-derived LST data can only be obtained under cloud-free conditions and because there are errors and calibration inconsistencies associated with many of these data. Owing to the great potential in using LST as a proxy for regional-scale surface mass balance, there is a need for more *in situ* validation work over ice-covered regions in order to assess the accuracy of satellite-derived LST over Arctic glaciers and ice caps.

7.5. Ice dynamics and iceberg calving

- Iceberg calving from ocean-terminating glaciers can be an important process of mass loss. Rates of calving loss can change rapidly for reasons that are influenced, but not directly controlled, by climate. These changes can generate anomalously high rates of mass loss, but not anomalously high rates of mass gain.
- Possible controls on calving rates include changes in lubrication from seasonal meltwater input, changes in resistance to flow from thinning and flotation, loss of contact with a stabilizing moraine, and break-up of floating ice tongues, as well as from changes in upstream dynamics, including surging.
- A tidewater glacier cycle is recognized. It involves prolonged periods of slow (~1000 years) glacier advance into the ocean, alternating with shorter periods (around decades) of accelerated glacier flow and retreat.

- Although measurements of iceberg calving fluxes are now available for many regions of the Arctic, the ability to model the processes that determine these fluxes is rudimentary. There is currently no ability to predict how calving fluxes will evolve in the future.
- The character of a glacier terminus and the icebergs produced are linked. Small, hard-to-detect bergs are produced mainly by grounded termini and glaciers in rapid retreat, while large tabular bergs come from floating ice tongues and ice shelves.
- The largest remaining ice shelves in the Arctic, fringe the northern coastline of Ellesmere Island. These entered a new phase of break-up after 2000. As a result, several fjords in the region are now free of a permanent ice cover for the first time in at least 3000 years.

7.5.1. Overview

Glaciers that terminate in the ocean or in lakes lose mass by iceberg calving as well as by melting and sublimation, and iceberg calving can be the dominant mass-loss term for many glaciers and ice caps in the Arctic. Rates of iceberg calving from marine-ending glaciers (referred to as 'tidewater glaciers' if their termini are grounded below sea level) are modulated by stress and flow conditions near the glacier's terminus, which will vary as the glacier changes shape and volume in response to climate change. Rates of calving can also change in response to internally controlled processes that alter stresses and may be entirely independent of climate. In either case, glacier mass loss through calving may be sudden and rapid, and may be far greater in magnitude than losses due to climatically controlled mass balance. Rapid dynamic changes can cause anomalously rapid mass loss, but not anomalously rapid mass gain, because alterations in flow can force fast ablation (by moving ice either to lower and warmer elevations or into the ocean), while accumulation rates are limited by precipitation.

7.5.1.1. Hydrology and basal sliding

Glaciers that slide over their beds typically exhibit significantly larger velocities and fluxes than glaciers that flow by internal (ice) deformation alone. Sliding is facilitated by the presence of water and deformable sediments at the ice/bed interface, although the relationship between water input and sliding speed is highly non-linear. Warmer surface conditions that increase surface melting in summer will accelerate glacier flow, but there is no simple relationship between the amount of warming and the magnitude of the flow acceleration. Higher water fluxes through subglacial drainage systems may result in greater enlargement of individual drainage channels, lower subglacial water pressures, and less flow enhancement than is generated by smaller fluxes. Thus, it is by no means clear that climate warming will, in the long term, result in more rapid glacier flow. As on Greenland, however, supraglacial lakes often form on Arctic glaciers in summer (Figure 7.16), and may drain suddenly when water-pressure induced fracturing creates drainage connections between the glacier surface and the glacier bed (Boon and Sharp, 2003). These lake drainage events may result in short-lived accelerations of glacier flow. As yet, however, there is no general rule to link increased sliding speed with increased englacial or subglacial water input, and no validated, reliable numerical model for basal sliding exists (Fountain and Walder, 1998; Hooke, 1998; Marshall et al., 2002).

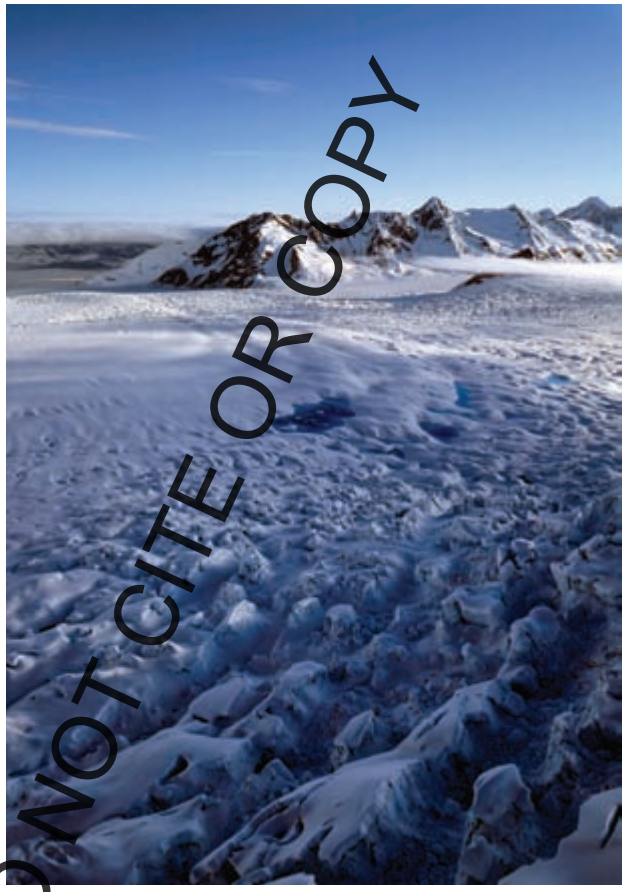


Figure 7.16. Supraglacial lakes forming on the surface of Columbia Glacier, Alaska, during the melt season. Source: W. Tad Pfeffer, Institute for Arctic and Alpine Research, University of Colorado

7.5.1.2. Glacier surges

Glacier surges are episodes of anomalously rapid motion (about 10 to 100 times 'normal' or non-surging speeds) that last for months to a few years and alternate with longer periods (decades to centuries) of normal, or quiescent, glacier behavior. Glacier velocities during the active phase of the surge cycle may reach 100 m per day and rates of terminus advance may approach 1 km per month. Glacier surges are not a direct response to climate forcing, but surging can alter a glacier's sensitivity to climate, especially in the immediate post-surge phase when its accumulation area is anomalously depleted (Dowdeswell et al., 1995).

7.5.1.3. Calving and marine-ending outlet glacier instability

'Marine-ending' or 'ocean-terminating' glaciers reach the ocean shore with sufficient flux to maintain an ice tongue that is either grounded in water shallower than the tongue's flotation depth or may be floating under polar conditions (subfreezing and, therefore, strong when in tension). Floating tongues can also form during periods of rapid terminus retreat.

The term 'tidewater glacier' is often used interchangeably with these labels, although it was originally restricted to glaciers with grounded termini. Marine-ending glaciers can undergo a periodic growth/shrinkage instability with long (roughly centuries) periods of slower motion and slow advance alternating with shorter periods (decades) of rapid (~5 to 10 km/y) motion, high calving flux, and rapid (~1 km/y) retreat (Meier and Post, 1987) (Figure 7.17). This is significant in that during rapid retreat a glacier can lose mass into the ocean far faster (by loss of its marine-grounded tongue and drawdown of its source basin by rapid flow) than is possible by direct surface mass balance forcing alone.

In its advancing phase, a tidewater terminus is stabilized in part by back stress against a moraine at its advancing margin. If it retreats from the moraine, a tidewater glacier in an advanced position can switch abruptly (within years) into an unstable phase of rapid flow and calving accompanied by thinning in its ocean-grounded reach that results in terminus retreat. Once initiated, rapid flow and retreat appear to be generally irreversible and continue until the glacier's terminus retreats inland to a point where the water depth provides no significant buoyancy force. The causes of unstable retreat are not completely understood, but involve high basal water pressures tied to the depth of the glacier bed below sea level, rapid sliding, vertical thinning, and positive feedback between thinning and rapid flow. The processes governing the size and rate of production of icebergs are extremely poorly known. Climate-induced thinning can play a critical role in initiating unstable retreat, but once initiated, unstable retreat appears to be modulated by channel geometry and englacial and basal hydrology and is essentially independent of climate and surface mass balance. Melting of the submarine faces of grounded tidewater glaciers can be rapid (Motyka et al., 2003) and can result in undercutting of the terminal ice cliff with consequence for iceberg production and terminus stability (Rignot et al., 2010; Straneo et al., 2010).

7.5.1.4. Lake-terminating glaciers

The total area of lake-terminating calving glaciers is much smaller than the area of marine-terminating glaciers, and the dynamics of their calving appear to be significantly different. Thinning rates on lake-calving glaciers are among the highest observed anywhere (e.g., -3.0 m/y basin-averaged thinning at Yakutat Glacier, Alaska) (Larsen et al., 2007).

7.5.1.5. Ice shelf / ice tongue break-up

Stable floating ice tongues (or ice shelves) are held in place by stabilizing stresses supplied by lateral attachments to valley walls or embayment margins (and, to a lesser degree, back stress from sea ice and confined iceberg fragments: *sikkusaq*), rather than by basal drag or back stress on the glacier terminus (Dowdeswell et al., 2000; Reeh et al., 1999). Submarine melt can be a significant source of mass loss from floating tongues (Mayer et al., 2000). Spreading and thinning of floating ice produces small tensile stresses, crevassing, and calving of icebergs. Under warming conditions, a floating ice tongue can be destabilized by (i) increased submarine melt, (ii) increased internal ice temperature, and (iii) percolation of surface

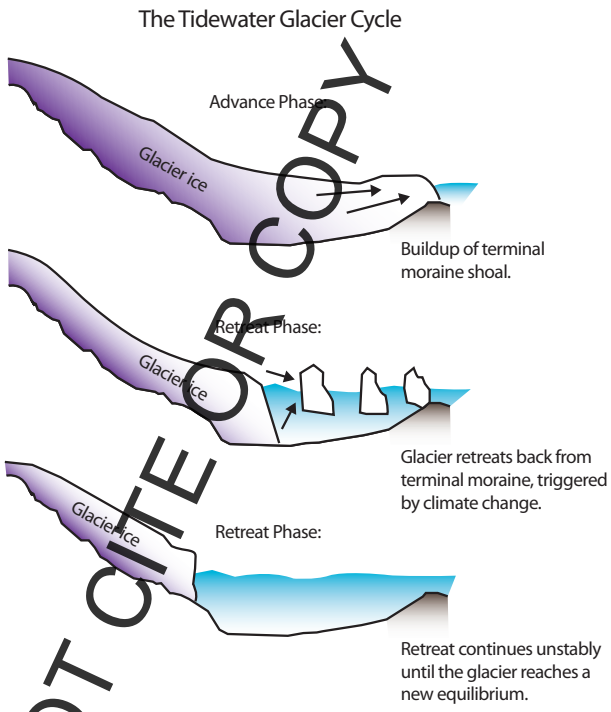


Figure 7.17. Schematic diagram showing the typical advance-retreat cycle of a tidewater glacier. Source: after Meier and Post (1987).

meltwater into open crevasses and fractures. Loss of an ice shelf or floating tongue and the associated back stress on the feeding glacier(s) affect the force balance on the ice upstream from the grounding line and in general cause an acceleration of the grounded ice and increased ice discharge to the ocean. The loss of the floating tongue has no influence on sea level, but increased discharge from the grounded part of the glacier contributes to eustatic sea-level rise.

7.5.2. Measurement methods

Glacier dynamic changes are manifested principally in changes in speed, surface elevation, ice discharge, and iceberg calving. Consequently, surface observations to track moving features or measure surface altitude are critical to understanding the processes involved. Glacier velocity is determined from repeat measurements of the position of markers on the glacier surface made by optical survey (Harper et al., 2007), GPS (Howat et al., 2008), and by feature tracking using aerial photography (Krimmel, 2001) and ground-based photography (O'Neel et al., 2005). Velocities are also measured remotely by manual and automated feature tracking applied to a variety of visual satellite imagery (Scambos et al., 1992; Joughin, 2002; Dowdeswell and Benham, 2003) and by radar interferometry and speckle tracking (Joughin et al., 1999). Elevation changes are measured by photogrammetry, by repeat altimetry from airborne or spaceborne platforms (Arendt et al., 2002, 2006; Larsen et al., 2007; Howat et al., 2008), and to a lesser degree by ground-based GPS and optical surveys.

The terminus advance/retreat rate ($\partial x/\partial t$) is the difference between the calving rate (u_c) and the ice flow speed at the terminus (u_T): ($\partial x/\partial t = u_c - u_T$). Iceberg calving events are episodic and very short in duration, and the volume of individual icebergs is hard to determine from imagery. The

calving rate is therefore difficult to measure directly, and it is typically calculated as $u_c = \partial x/dt + u_T$ (O'Neel et al., 2005), but this is a weak approach in predictive models where $\partial x/\partial t$ is often the variable being sought. New methods are being developed to use seismic records to identify individual calving events and estimate iceberg discharge (O'Neel and Pfeffer, 2007).

7.5.3. Calving and surging glaciers in the Arctic

Surging and marine calving glaciers are found throughout the Arctic, although there are currently none in Norway or Sweden. Table 7.2 summarizes recent estimates of mass loss by calving from major regions and/or major ice masses within the Arctic (but note that no estimates are available for some regions and ice masses).

7.5.3.1. Alaska and western Canada

The greatest concentrations of glacier area in Alaska and northwestern Canada lie along the coast of the Gulf of Alaska, where land-terminating glaciers on the interior side drain predominantly into the Matanuska, Copper, and Yukon Rivers, while seaward-facing glaciers drain into, or terminate in, the Gulf of Alaska.

7.5.3.1.1. Calving

A significant but undetermined fraction of the area of Gulf of Alaska glaciers is drained through 54 calving tidewater outlets. No estimates have been made of calving flux for the entire Gulf, but a 1982 study (Brown et al., 1982) compared geometry and calving speed for 13 of the largest tidewater glaciers and found a total flux of 7.2 Gt/y draining from an aggregate basin area of 8138 km^2, estimated for various periods within the interval 1942 to 1979. No more recent or more complete assessment has been made.

In the past 200 years, the 54 tidewater glaciers terminating in the Gulf of Alaska all retreated from extended stable positions. Columbia Glacier was the last to start retreating and as of 2009 had retreated roughly two-thirds of the way up the 30-km long fjord in which it was grounded up to about 1982. The majority of Gulf of Alaska tidewater glaciers are now either stable in retreated positions or slowly advancing (e.g., Hubbard and Mears Glaciers). Columbia Glacier is undergoing the most rapid change, calving an average 3.3 Gt/y of ice (1982 to 2001), with a maximum annual calving flux of 6.6 Gt in 2001 (O'Neel et al., 2005), and thinning in the terminus region by 20 m/y. GRACE gravity measurements suggest that the Gulf of Alaska glaciers were losing mass at a rate of 84 ± 5 Gt/y

Table 7.2. Estimated mass losses by iceberg calving of mountain glaciers and ice caps. Specific balances due to calving are derived by dividing the calving mass loss by the total area of the glacier.

		Period	Area (km^2)	Calving loss (Gt/y)	Specific balance (m/y w.e.)	Reference
Alaska						
	Columbia Glacier	1982 – 2001	1000	3.3	-3.3	O'Neel et al., 2005
Canadian High Arctic						
	Devon Ice Cap	1960 – 1999	14 000	0.53±0.12	-0.04	Burgess et al., 2005
	Prince of Wales Icefield	1963 – 2003	19 300	1.9 ±0.02	-0.10±0.00	Mair et al., 2009
	Agassiz Ice Cap	1999 – 2002	19 500	0.60±0.14	-0.03±0.01	Williamson et al., 2008
	W Grant Ice Cap	1999 – 2003	2 000	~0.23	~ -0.12	Williamson et al., 2008
Russian Arctic						
	Academy of Sciences Ice Cap	1995	5 575	0.59	-0.10	Dowdeswell et al., 2002
	Franz-Josef Land	1929 – 1959	13 700	2.1	-0.15	Govorukha, 1989
		1952 – 2001		4.3	-0.31	Glazovsky and Macheret, 2006
	Novaya Zemlya	1930 – 1960	23 600	1.8	-0.08	Govorukha, 1989
		1952 – 2001		1.4	-0.06	Glazovsky and Macheret, 2006
	Severnaya Zemlya	1929 – 1972	19 400	0.45	-0.02	Govorukha, 1989
		1952 – 2001		0.7	-0.04	Glazovsky and Macheret, 2006
Svalbard						
	All glaciers	1968 – 1998	36 600	4±1	-0.1±0.0	Hagen et al., 2003b
		2000 – 2006	36 600	6.75±1.7	-0.18±0.05	Błaszczyk et al., 2009
	Austfonna	[b]	8 000	2.5±0.5	-0.31±0.06	Dowdeswell et al., 2008

[a] Calving estimates only from Otto Glacier which is likely the majority of the mass loss by calving from the ice cap; [b] a combination of various data sets from the period 1973 to 2001.

in the period 2003 to 2007 (Luthcke et al., 2008). A study of the southeastern (panhandle) Gulf of Alaska sub-region (not including Columbia Glacier) found a loss of 16.7 ± 4.4 Gt/y from 14 580 km^2. Two-thirds of this was from calving glaciers, although the fraction of this volume loss that was attributable to calving (as opposed to surface melting) is unknown (Larsen et al., 2007).

Here, and elsewhere, episodic discharges from individual glaciers can exceed the aggregate discharge from the rest of the region, making averages over time difficult to interpret or extrapolate.

7.5.3.1.2. Surging

Surging glaciers are abundant in the central Alaska Range (Denali massif), eastern Alaska Range, and throughout the St. Elias mountains, where 'Alaskan-type' surging was defined (Figure 7.18). Bering Glacier, in this region, is the largest glacier (and largest surging glacier) in Alaska (Shuchman and Josberger, 2010). Its last major surge took place in 1995 to 1996, although there was a small surge in 2009 to 2010. Bering Glacier terminates in a large proglacial lake at sea level, the size of which changes dramatically as the glacier advances and retreats. Some surging glaciers in the St. Elias Range terminate in the ocean, allowing surge behavior to influence calving discharge. No surging glaciers are known in the Chugach or Brooks Ranges.

7.5.3.2. Arctic Canada

7.5.3.2.1. Calving

There is no comprehensive summation of calving flux or inventory of calving glaciers for the Canadian Arctic Archipelago, but estimates of calving fluxes for some of the ice caps exist for mixed time periods in the interval 1960 to 2003, totaling 3.0 Gt/y (Devon Ice Cap, 0.5 Gt/y; Agassiz Ice Cap, 0.6 Gt/y; Otto Glacier (western Grant ice complex), 0.2 Gt/y; Prince of Wales Icefield, 1.7 Gt/y). As in Alaska, individual glaciers can dominate the regional flux estimates.

7.5.3.2.2. Surging

At least 51 surge-type glaciers have been identified in the Queen Elizabeth Islands (Copland et al., 2003). Of these, 15 were surging in 1959/1960 and/or 1999/2000. The largest recorded surge-type advances (4 to 7 km) were observed on Axel Heiberg Island and the Manson Icefield (Ellesmere Island). The active phase of the surge cycle of these glaciers may last more than a decade (Mueller, 1969; Copland et al., 2003), typical of 'Svalbard-type' surging (Mueller, 1969; Murray et al., 2003). Glacier velocities during the active phase of the surge cycle may range from several hundred metres to over 1 km per year. Because many of the known surge-type glaciers terminate in the ocean, surging may cause large short-term changes in the loss of mass to the oceans by iceberg calving. Little is known about the duration of the quiescent phase of the surge cycle in this region owing to the short period of observations.

Figure 7.18. Hubbard Glacier, which drains into Russell Fjord in the Yakutat region of southern Alaska, is a surging tidewater glacier. Source: Anthony Arendt, University of Alaska.

7.5.3.3. Greenland (excluding the Greenland Ice Sheet)

Little is known about the history of area and volume changes of the roughly 48 600 km^2 of glaciers and ice caps that surround the Greenland Ice Sheet, or about their mass balance, calving discharge, or the fraction of their area drained through marine-terminating outlets. Mass balance estimates for these glaciers have been made (Dyurgerov and Meier, 2005) using correlated mass balance records from Svalbard and the Canadian Arctic Archipelago. Net losses from 1992 to 2002 are estimated to be about 18 Gt/y, but the fraction due to calving is completely unknown.

7.5.3.4. Iceland

In Iceland, 26 surge-type glaciers and 80 surge advances have been identified (Björnsson et al., 2003), and surges account for a significant fraction of the mass transported by major outlet glaciers of all the main ice caps. Surges affected 38% of the area of Vatnajökull in the 1990s, and accounted for about 25% of the mass transfer from the accumulation zone of the ice cap to the ablation zone (Björnsson and Pálsson, 2008). Calving losses are unknown but probably not significant.

7.5.3.5. Svalbard

Glacier geometry, flow velocity, and front position changes of Svalbard glaciers were determined from ASTER images acquired in 2000 to 2006 (Błaszczyk et al., 2009). A total of 163 grounded tidewater glaciers drain an area of around 21 200 km^2 (more than 60% of the total glacier area of Svalbard) through a combined calving terminus length of 860 km.

7.5.3.5.1. Calving

Mass loss due to calving from the whole archipelago was initially estimated to be about 4 Gt/y for the period 1968 to 1998 (Hagen et al., 2003b), but a more recent estimate for the period 2000 to 2006 is 5.0 - 8.4 Gt/y, with a mean value of 6.75 ± 1.7 Gt/y (Błaszczyk et al., 2009). The mean retreat rate (dx/dt) of the calving glaciers is about 30 m/y (~2.1 Gt/y), and terminus retreat ($\dot{M}_L$) accounts for about 30% of the total calving flux from the archipelago. Calving constitutes about 20% of the overall mass loss from Svalbard glaciers (Błaszczyk et al., 2009).

Dowdeswell et al. (2008) estimated the calving flux from

Austfonna to be 2.5 ± 0.5 Gt/y from an aggregate terminus length of about 230 km. This represents 30% to 40% of the annual ablation from this ice cap. Terminus retreat reduced the ice cap area by about 10 km²/y during the period 1991 to 2001 and accounted for about 50% of the calving loss.

7.5.3.5.2. Surging

Surge-type glaciers are very common in Svalbard and comprise a range of glacier types from small land-terminating cirque glaciers to large calving outlets. Although the overall fraction of surge-type glaciers is unknown, estimates range from 13% to 90% (Liestøl, 1969; Dowdeswell et al., 1991; Lefauconnier and Hagen, 1991; Hamilton and Dowdeswell, 1996; Jiskoot et al., 1998). Surging of marine-grounded outlets can interact with calving dynamics. The 1250-km² Hinlopenbreen calved about 2 km³ of icebergs in a single year during a surge (Liestøl, 1973). Bråsvellbreen (1100 km²) advanced up to 20 km along its 30-km marine-grounded margin during a surge in 1936 to 1938 (Schytt, 1969). In the inner part of Storfjorden, east Spitsbergen, Negribreen advanced about 12 km in less than a year (about 35 m/d) along a 15-km long section of its front during a surge in 1935 to 1936 (Liestøl, 1969). Recently, the tidewater glacier Nathorstbreen advanced about 7.6 km or about 20 m/d from September 2008 to September 2009 (Sund and Eiken, 2010).

7.5.3.6. Russian Arctic islands

7.5.3.6.1. Calving

Govorukha (1989) estimated calving fluxes of 1.8 Gt/y from Novaya Zemlya (1930 to 1960), 0.45 Gt/y from Severnaya Zemlya (1929 to 1972), and 2.1 Gt/y from Franz Josef Land during the period 1929 to 1959 (a total of 4.35 Gt/y). Using measurements of terminus ice thickness derived by radio echo sounding, ice surface velocities at glacier termini derived by radar interferometry, and measurements of terminus length and terminus advance/retreat, Glazovsky and Macheret (2006) estimated calving fluxes from the Russian Arctic islands to be 6.4 Gt/y for the period 1952 to 2001 (Novaya Zemlya = 1.4 Gt/y; Severnaya Zemlya = 0.7 Gt/y; Franz Josef Land = 4.3 Gt/y). Losses due to terminus retreat were estimated to be about 1.35 Gt/y, or roughly 21% of the total calving flux. It is not possible to determine the significance of the difference between the two calving flux estimates, so it should not be taken to imply a temporal trend in calving fluxes.

7.5.3.6.2. Surging

Surge-type glaciers are rare in the Russian Arctic Archipelago (Dowdeswell and Williams, 1997). None are known to exist in Franz Josef Land, while until recently only three had been identified in Novaya Zemlya and two in Severnaya Zemlya. However, a new analysis based on high-resolution satellite imagery identified 32 potential surge-type glaciers on Novaya Zemlya, representing 18% of the total glacier area (Grant et al., 2009). These glaciers are generally relatively long (median length 18.5 km), large (median area 106.8 km²) outlet glaciers that terminate in water. They occupy large, complex catchment areas that may be receiving increased precipitation from the Barents Sea.

7.5.4. Recent ice shelf break-up events

Most of the largest floating ice shelves in the Arctic, fringing the northern coast of Ellesmere Island, are not predominantly glacier-fed, but form mainly by basal accretion of sea ice and surface accumulation of precipitation (Jeffries, 1992). They developed around 5500 to 3000 years BP (England et al., 2008) and have lost over 90% of their area during the 20th century, with only four major ice shelves now remaining. Much of the area loss occurred in the 1950s (Koenig et al., 1952) and 1960s (Hattersley-Smith, 1963, 1967). A new phase of disintegration began in 2000, with renewed fracturing of the Ward Hunt Ice Shelf (Mueller et al., 2003) and continued in 2005 with the abrupt loss of almost all of the Ayles Ice Shelf and major calving from the Petersen Ice Shelf (Copland et al., 2007). In summer 2008, the total ice shelf area in Arctic Canada decreased further (by 23% relative to 2007; Mueller et al., 2008) due to the loss of the entire Markham Ice Shelf and 60% of the area of the Serson Ice Shelf. As a result, several fjords on the northern coast of Ellesmere Island are now ice-free for the first time in 3000 to 5500 years (England et al., 2008). In 2010, large new fractures in the Ward Hunt Ice Shelf were first detected in Radarsat-2 images from 7 and 14 August. A MODIS image from 18 August shows that breakup of the eastern part of the ice shelf was underway (Figure 7.19), and some 65-70 km² of the shelf had been lost by the end of August. Meanwhile, fragments of the ice islands that calved from the Ayles, Serson, Peterson, Ward Hunt and Markham ice shelves in 2005 and 2008 have drifted into the Canada Basin and the Sverdrup and Queen Elizabeth Islands and are beginning to enter the Northwest Passage via Barrow Strait.

Causes of ice shelf break-up appear to include rising summer air temperature and the loss of semi-permanent landfast sea ice along the northern coast of Ellesmere Island, although strong southerly winds seem to play an important role in individual break-up events. Ice shelf break-up events produce large 'ice islands' which may become a hazard to shipping and offshore resource exploration if they drift westward into the Beaufort Sea. They can also result in the drainage of epishelf lakes and the loss of their unique ecosystems (Veillette et al., 2008).

As with floating ice tongues, the loss of ice shelves does not directly influence sea-level rise. In the case of the Ellesmere ice

Figure 7.19. MODIS image from 18 August 2010 showing the Ward Hunt Ice Shelf, northern Ellesmere Island, and the extensive fracturing that had developed in the ice shelf to the east and south of Ward Hunt Island. Source: Canadian Ice Service.

shelves, because most of them do not interact with terrestrial glaciers, there is no indirect effect on sea level.

The Matusevich Ice Shelf in Severnaya Zemlya, the only major ice shelf in the Eurasian Arctic, has also undergone periodic calving events during roughly the past 70 years (Williams and Dowdeswell, 2001).

7.5.5. Iceberg characteristics and relationship to ice dynamics

There are certain relationships between iceberg characteristics (size, number, morphological type) and parent glacier type and behavior. These relationships can be broadly described as follows (Dowdeswell, 1989; Dowdeswell et al., 1992):

- The largest tabular icebergs (large horizontal dimension relative to depth) are calved from floating glacier tongues and ice shelves. Horizontal sizes reach hundreds of metres to kilometres. The number of individual icebergs calved each year is usually small.
- Fast-flowing tidewater glaciers that are grounded have high rates of discharge, but often produce icebergs of relatively small size (tens to hundreds of metres) and irregular shape that tend to break up quickly into still smaller bergs and bergy bits (Dowdeswell and Forsberg, 1992) (Figure 7.20). Slower-flowing glaciers calve fewer icebergs, but their sizes may be larger than those of icebergs calved from more active glaciers.
- Surging glaciers and tidewater glaciers in rapid retreat tend to produce a very large number of small icebergs (10 to 50 m in length) with irregular shapes (Schytt, 1969). If such glaciers start to float, however, the mode of calving may change to one in which episodic flow-perpendicular rift formation and propagation result in the release of a few very large icebergs (Walter et al., 2010).

The systematic classification of iceberg types made by Dowdeswell for glaciers of Spitsbergen can be applied to all sources of icebergs in the Arctic. It is supported by descriptions of icebergs made at different times near Spitsbergen, Franz Josef Land, and Novaya Zemlya (Sandford, 1955; Kubyshkin et al., 2009), and by observations in the northern Barents Sea in 2003 to 2007. For example, tidewater outlet glaciers in Russkaya Gavan Bay and Inostrantsev Bay in northern Novaya Zemlya have strong fracturing of the front and produce a moderate quantity of small- and medium-sized icebergs of irregular shape (Buzin and Glazovsky, 2005). Floating margins on some of the larger ice caps of Franz Josef Land produce large tabular icebergs (Dowdeswell et al., 1994; Zubakin et al., 2007).

7.5.6. Controls on calving fluxes

Although the detailed physics of iceberg calving are poorly understood, certain controlling factors can be identified. The tensile strength of ice is strongly dependent on temperature, and temperate ice (ice at its pressure melting point, where water and ice exist in thermal equilibrium) has essentially no tensile strength (Schulson, 1999). Along-flow extension or compression of ice influences the opening or closing of transverse crevasses, but no clear relationship between calving rate and along-flow strain rate has been demonstrated

Figure 7.20. Small iceberg off the coast of Franz Josef Land. Note the debris-rich layers within the iceberg. Source: Roman Romanov, [affiliation].

observationally on meaningful timescales (Alley et al., 2008). A statistical relationship between water depth at grounded glacier termini and calving rate has been demonstrated (Brown et al., 1982), but this relationship is purely empirical and possibly applicable only to calving margins near steady-state. Sea ice, floating brash ice, and iceberg debris can exert substantial restraining forces (or back stresses) that are transmitted to the fjord walls and inhibit calving (Amundson et al., 2010). The presence or absence of a moraine shoal at the terminus providing resistive back stress against extension of ice in the direction of flow has a strong influence on calving rate (Meier and Post, 1987; Fischer and Powell, 1998; Nick et al., 2007). However, it is unclear whether the detachment of a glacier terminus from its terminal moraine is a cause or a consequence of thinning and retreat.

7.5.7. State of theory and modeling

A number of attempts have been made to formulate calving models, but there is still no universal theoretical framework that can explain the full suite of available observations. Using data from twelve Alaskan glaciers, Brown et al. (1982) proposed an empirical linear relationship between the calving rate and the water depth at the terminus. However, the correlation is not broadly applicable because there is no obvious physical control that can explain any depth dependency of the calving rate (Benn et al., 2007). A drawback to such empirical models is that they are often based on data from glaciers that are either in or near a steady-state (Van der Veen, 1996). Indeed, based on observations at Columbia Glacier in Alaska, Meier and Post (1987) suggested that a linear calving relationship might not be valid during periods of rapid advance or retreat. Various approaches to a mechanically based process model of calving have been pursued since the mid-1960s, based on several different lines of analysis. These include, most notably, an elastic bending beam model developed by Reeh (1968), analyses of crevasse penetration with and without water (Nye, 1957; Weertman, 1973), and analyses based on linear fracture mechanics (Van der Veen, 1998a,b). Other observations and analyses include work by Holdsworth (1969) and Hughes (1992).

Based on observations from Columbia Glacier, Van der

Veen (1996) proposed a flotation model, wherein the terminus of the glacier retreats to the position where the ice thickness exceeds the flotation thickness by some threshold value, H_0. On Columbia Glacier, H_0 was determined to be about 50 m. Vieli et al. (2001) pointed out that for smaller glaciers (water depths below 150 m and flow velocities less than 200 m/y), frontal cliff heights do not usually exceed 50 m, so the height above flotation will be much less than was suggested for Columbia Glacier. As a result, they proposed a modified flotation criterion, with the minimum height above flotation, H_0, replaced by a fraction, q, of the flotation thickness at the terminus. Benn et al. (2007) noted that one significant weakness of flotation models is that they do not allow floating termini to form, thus limiting their use to certain types of calving margins. As a result, they proposed a new criterion based on a simple model of crevasse formation, in which the calving margin is located where crevasse depth is equal to the glacier surface height above sea level, and the determination of the calving rate is reduced to determining ice thickness, velocity distribution, and changes in thickness and velocity through time.

Hanson and Hooke (2000) treated calving as a multivariate problem, using water depth, longitudinal strain rate, and ice temperature as the primary factors controlling the calving rate. Nick and Oerlemans (2006) compared the water depth and flotation models using synthetic simplified glacier geometries and found the latter to be superior, but they failed to simulate a full cycle of glacier length variations when the glacier terminates in very deep water. Nick et al. (2007) elaborated on their model by including a simple sediment transport scheme in the ice flow model and concluded that sedimentation at the glacier front needs to be included for the glacier to advance into deepening water.

Most calving models presented to date are either wholly empirical or based on mechanics. Other englacial processes, including glacial hydraulics, have not been incorporated, although there is evidence that water flow along cracks may play an important role in the calving process (O'Neel and Pfeffer, 2007).

No validated, broadly applicable, and robust calving model exists at this point. The calving relationships and models proposed so far have been tailored and applied to specific glaciers and data sets, and the proposed models tend to require data that are not readily available on regional scales. Larger models, either focused glacier/ice shear models or comprehensive coupled GCMs, tend to include calving only in an extremely simplistic fashion (e.g., specifying a fixed calving boundary location) or do not treat calving at all. Hence, there is a pressing need for robust calving formulations of intermediate complexity that are capable of being applied in a wide variety of settings and for regional-scale mass change predictions. There is also a need for the basic data required to constrain models based on such a formulation.

7.5.8. Requirements for improved predictions of calving fluxes

Because calving can deplete terrestrial ice reservoirs very rapidly, calving dynamics are a key element in understanding and predicting the mass balance and sea level rise contributions of all glaciers, ice caps, and ice sheets with calving margins. Knowledge of present-day calving rates is extremely limited owing to a lack of fundamental observations. Since changes in flow dynamics and calving are not controlled directly by externally observed variables (such as climate), robust predictions depend on an understanding of glacier physics that is still lacking and on observations that can be difficult to make. The most fundamental observations needed to better constrain potential calving losses are (i) more complete observations of calving rate and marine-ending terminus changes, and (ii) radio echo sounding of ice thicknesses and basal topography to identify marine-grounded outlets and the marine-grounded fractions of glacier and ice cap areas. Without these measurements, the potential for dynamically forced calving losses cannot be constrained let alone predicted.

7.6. Projections of Arctic glacier changes

- Most attempts to simulate the response of Arctic glaciers to future climate change involve evaluation of the response of the surface mass balance to prescribed changes in climate (usually air temperature and precipitation); simulations generally do not include changes in glacier dynamics or mass loss by iceberg calving. Climate change can be imposed as either step changes from an initial state or as transient changes over some period of time derived from climate models.
- Projections for individual glaciers throughout the Arctic show substantial mass loss or even disappearance of smaller glaciers by the end of the 21st century in response to imposed temperature and precipitation scenarios.
- Very few studies have attempted to model the response of Arctic glaciers on larger scales. A new simulation of the surface mass balance of the global population of mountain glaciers and ice caps to 2100 uses a temperature index mass balance model driven by downscaled output from ten GCMs, all forced by the IPCC A1B emissions scenario. For Arctic glaciers, projected volume loss due to surface mass balance ranges between 0.049 to 0.131 m sea-level equivalent, or 12% to 32% of their current volume by 2100 depending on the choice of GCM. Most models suggest that the largest mass losses are from glaciers in Alaska and Arctic Canada.

7.6.1. Downscaling climate model projections

General circulation models (Solomon et al., 2007) driven by standardized IPCC emissions scenarios (B1, A2, A1B; see Chapter 3) generally predict that warming and precipitation increases in the Arctic beyond 2030 will be larger than the global mean (Chapman and Walsh, 2007; Kattsov et al., 2007). Only part of the projected increases in precipitation will lead to increases in snow accumulation because the fraction of precipitation falling as rain will increase with the projected air temperature increases.

Direct use of data from GCMs for glacier mass balance projections is not currently feasible, owing to the generally large biases in model data on regional scales. Individual glaciers are typically much smaller than climate model grid boxes, and often occupy complex terrain that is only coarsely resolved in the model's underlying topography. Mass balance modeling is

particularly sensitive to biases and offsets in air temperature, which controls the energy available for melt and the snow-rain ratio of precipitation (and thus controls both ablation and accumulation). Hence, some form of downscaling is required to transfer global-scale climate information to local glacier scales prior to making glacier mass balance projections.

Statistical downscaling is often used, whereby statistical relationships are established between meteorological quantities determined at the GCM and the local scales. This can be done using either field observations or other suitable meteorological data such as data from climate re-analyses. Radić and Hock (2006) applied 'local scaling' to correct for the biases in climate model air temperature and precipitation for projecting the mass balance of Storglaciären until 2100. Downscaled temperature series were produced from GCMs and regional climate models (RCMs) by shifting the series by the averaged monthly differences between climate model and local-scale data (here ERA-40 re-analysis data) over a baseline period for which both GCM and local-scale data were available. Hence, the average seasonal cycle from ERA-40 was used as a reference by which seasonal cycles from the climate model could be 'corrected'. The results highlight the importance of including seasonally varying biases in the downscaling instead of assuming a constant bias throughout the year. Precipitation was scaled by the ratio of precipitation in the reference data set (here ERA-40) summed over the baseline period and the corresponding climate model precipitation sum. A drawback of statistical downscaling is the inherent assumption that the statistical relationships established over a baseline period will continue to hold in future climates.

An alternative approach to statistical downscaling is to use changes in GCM variables between a defined time slice in a GCM simulation of future climate and a baseline period corresponding to a period for which local climate data are available. These changes are then used to perturb the observed local climate data to drive a mass balance model and project future mass balance changes (e.g., Schneeberger et al., 2003). The changes in climate variables are often simply linearly interpolated between time slices to allow for transient simulations. Due to generally large interannual variability, however, projected climate variable changes can be sensitive to the choice of baseline period. This is especially true when large fluctuations occur around the baseline period or when the climate variable shows a trend during this period (Aðalgeirsdóttir et al., 2006).

Zhang et al. (2007a) used a dynamical downscaling approach, employing the high resolution regional model Polar MM5 driven by global atmospheric re-analysis to obtain temperature and precipitation data on a 10-km resolution grid to force a glacier mass balance model. The results of mass balance simulations using dynamically downscaled data and simulations based on observed temperature and precipitation data were in reasonably good agreement when calibration was used to minimize systematic biases in the MM5 downscaling.

7.6.2. Modeling mountain glacier and ice cap mass balance in the 21st century

While numerous studies have projected the response of individual glaciers to climate change, very few studies have attempted to project the response of Arctic glaciers on a regional scale. Zhang et al. (2007b) used dynamically downscaled daily maximum and minimum air temperatures and precipitation driven by the IPCC A1B emissions scenario of the CCSM3 climate model to force a temperature-index mass balance model for Hubbard Glacier (2460 km^2) and Bering Glacier (3630 km^2) in Alaska for the period 2010 to 2018. Both glaciers are projected to have increased accumulation, particularly on the upper reaches of the glaciers, and increased ablation, particularly on the lower parts. The modeled mass balance of Bering Glacier is projected to become more negative on average (-2.0 m w.e./y compared to -1.3 m w.e./y during the 1994 to 2004 baseline period) and the modeled mass balance of Hubbard Glacier, which was positive during the baseline period, is projected to decrease slightly (to 0.3 m w.e./y from 0.4 m w.e./y).

Hock et al. (2007a) modeled cumulative mass balance changes of the 3 km^2 Storglaciären in Sweden until 2100 in response to model-projected regional temperature changes. They predicted total changes ranging from -81 to -121 m w.e., depending on the choice of the mass balance model. A fully distributed energy balance melt model produced a greater mass balance response than simpler temperature-index models. Projections based on the IPCC B2 emissions scenario showed a 50% to 90% decrease in ice volume by 2100 (Radić and Hock, 2006) (Figure 7.21). The volume change projections vary by 40% of the initial ice volume for six different GCM inputs to the mass balance model, and by 10% depending on the details of the mass balance model used. This is in agreement with the conclusion of Oerlemans et al. (2005) that the uncertainties due to a simple representation of glacier processes are less than those associated with the output from GCMs.

Huss et al. (2008) modeled the complete disappearance of the 4.4 km^2 Laika Glacier in Arctic Canada by 2100 in response to temperature and precipitation trends derived from a climate

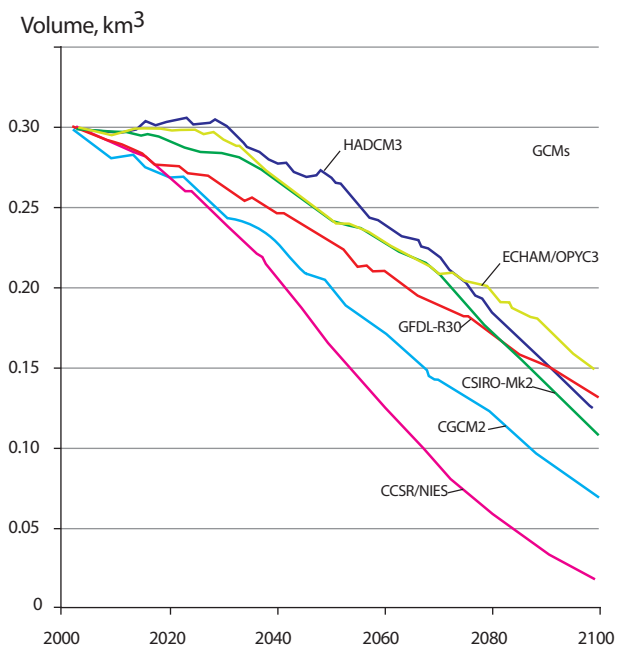


Figure 7.21. Volume projections for Storglaciären, Sweden, in the 21st century derived from six general circulation models (GCMs). Source: Radić and Hock (2006).

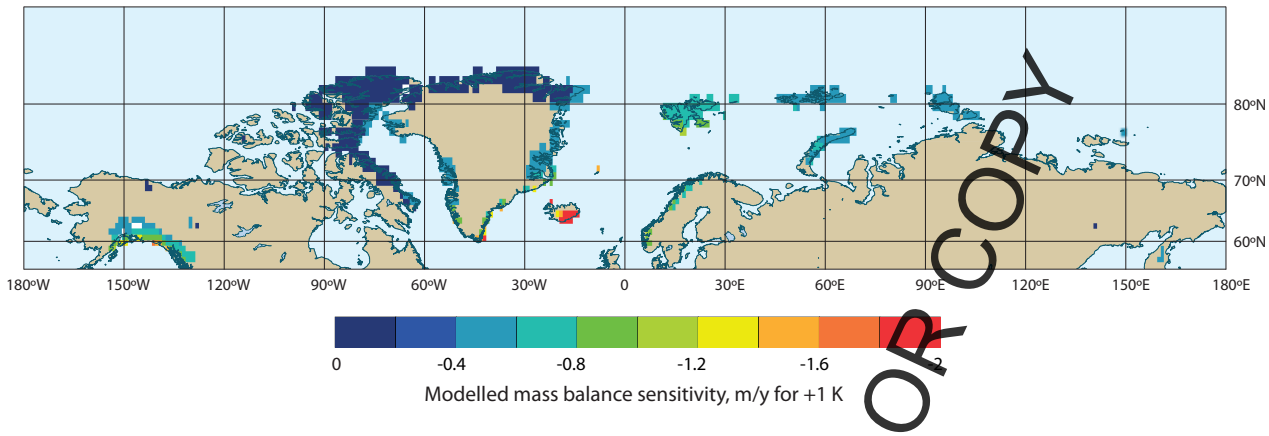


Figure 7.22. Modeled mass balance sensitivity (m w.e./y per +1 K), that is, the change in specific mass balance owing to a 1 K uniform warming. Source: Hock et al. (2009).

model forced by the IPCC A1B emissions scenario. Bassford et al. (2006) modeled complete wastage of Pioneer Ice Cap (~200 km^2) and Vavilov Ice Cap (~1770 km^2) on Severnaya Zemlya within about 370 and about 1080 years, respectively, by imposing a linear change in temperature and precipitation over the period 1990 to 2100 based on the IPCC IS92a 'business-as-usual' scenario, after which the climate was held constant. Although these ice caps have quite low static sensitivities of mass balance to changes in temperature and precipitation, they are nonetheless highly sensitive to long-term climate change owing to their hypsometry and location on relatively flat beds close to sea level.

Ananicheva and Krenke (2007) modeled mass changes of 17 glacier regions in northeastern Siberia and Kamchatka (1040 km^2) for the period 2040 to 2069 using climate projections from the ECHAM4 GCM driven by the IPCC A2 emissions scenario. The equilibrium-line altitude (ELA) on these glaciers is predicted to rise by 230 m in the northern parts and 500 m in the southern parts of northeastern Siberia. The upward shift is largest in Kamchatka. Based on known correlations between the ELA and the elevation of the glacier terminus, only 22% and 31% of the glacier areas are predicted to remain in northeastern Siberia and Kamchatka, respectively, in 2070.

On larger regional or global scales, mass changes have been projected using either an 'indirect' approach based on mass balance sensitivity, defined as the change in mass balance caused by an instantaneous change in a climatic variable such as air temperature or precipitation, or a 'direct' approach based on transient mass balance modeling. The indirect approach involves computation of sensitivities, often for each month individually ('seasonal sensitivity characteristic'; Oerlemans and Reichert, 2000), which are used to calculate the change in surface mass balance for given anomalies of temperature and precipitation. Mass balance sensitivities are determined for individual glaciers by energy balance or temperature-index modeling. Mass balance sensitivities to temperature computed on a 1° by 1° grid exhibit large spatial variations across the Arctic, with the largest sensitivities in Iceland, southern Greenland, and coastal Alaska, and the smallest sensitivities in Arctic Canada and northern Greenland (Hock et al., 2009; Figure 7.22).

Applying a regression-based temperature index model to 42 glaciers located north of 60° N, de Woul and Hock (2005) found that the mass balance sensitivity to a hypothetical temperature increase of +1 K ranged from -0.2 to -2.0 m/y. Sensitivity of the mass balance to a 10% increase in precipitation ranged from < +0.1 to +0.4 m/y. This offset the effect of a +1 K temperature increase by about 20% on average. Maritime glaciers had considerably higher mass balance sensitivities than continental glaciers. Results were extrapolated to the entire Arctic by assuming area-weighted means of sensitivity to be representative for different Arctic regions. The estimated total contribution to sea level from Arctic glaciers (including the Greenland Ice Sheet) in response to a +1 K warming was about +0.6 mm/y. This compares to an estimate of +0.53 mm/y derived using a sensitivity approach that suggested about equal contributions from the Greenland Ice Sheet and the Arctic glaciers (Oerlemans et al., 2005).

Radić and Hock (2011) computed global mass balance and volume changes for more than 120 000 mountain glaciers and roughly 2600 ice caps around the world until 2100 using an elevation-dependent temperature-index mass balance model driven by output from ten GCMs forced by the IPCC A1B emissions scenario. Future volume changes were up-scaled to all glaciers using a regionally differentiated approach. For the Arctic glaciers (including those in Greenland surrounding the ice sheet), the projected volume loss due to melt (ablation by calving is not included) ranges from 51 to 136 mm sea-level equivalent (SLE) (i.e., 13% to 36% reduction in their current volume by 2100) depending on the choice of GCM (Figure 7.23). The multi-model mean is 88 ± 28 mm SLE, or 22% ± 7% volume reduction where the uncertainty range is ±1 standard deviation. Multi-model mean volume reductions by 2100 (relative to initial volumes) vary considerably among regions, with the smallest values in Greenland (8% ± 5%) and the largest values in Svalbard (54% ± 15%); however, the spread among models for most regions is large (Figure 7.23). Most models show the largest sea level contributions coming from the Canadian Arctic and Alaska (including the Yukon), followed by the Russian Arctic and Svalbard. Contributions from Arctic Canada show a very large range from less than 10 mm SLE contribution to the largest projected contribution for any of seven Arctic regions, indicating a large spread in the projected temperature and precipitation fields among the GCMs for this region (Figure 7.24).

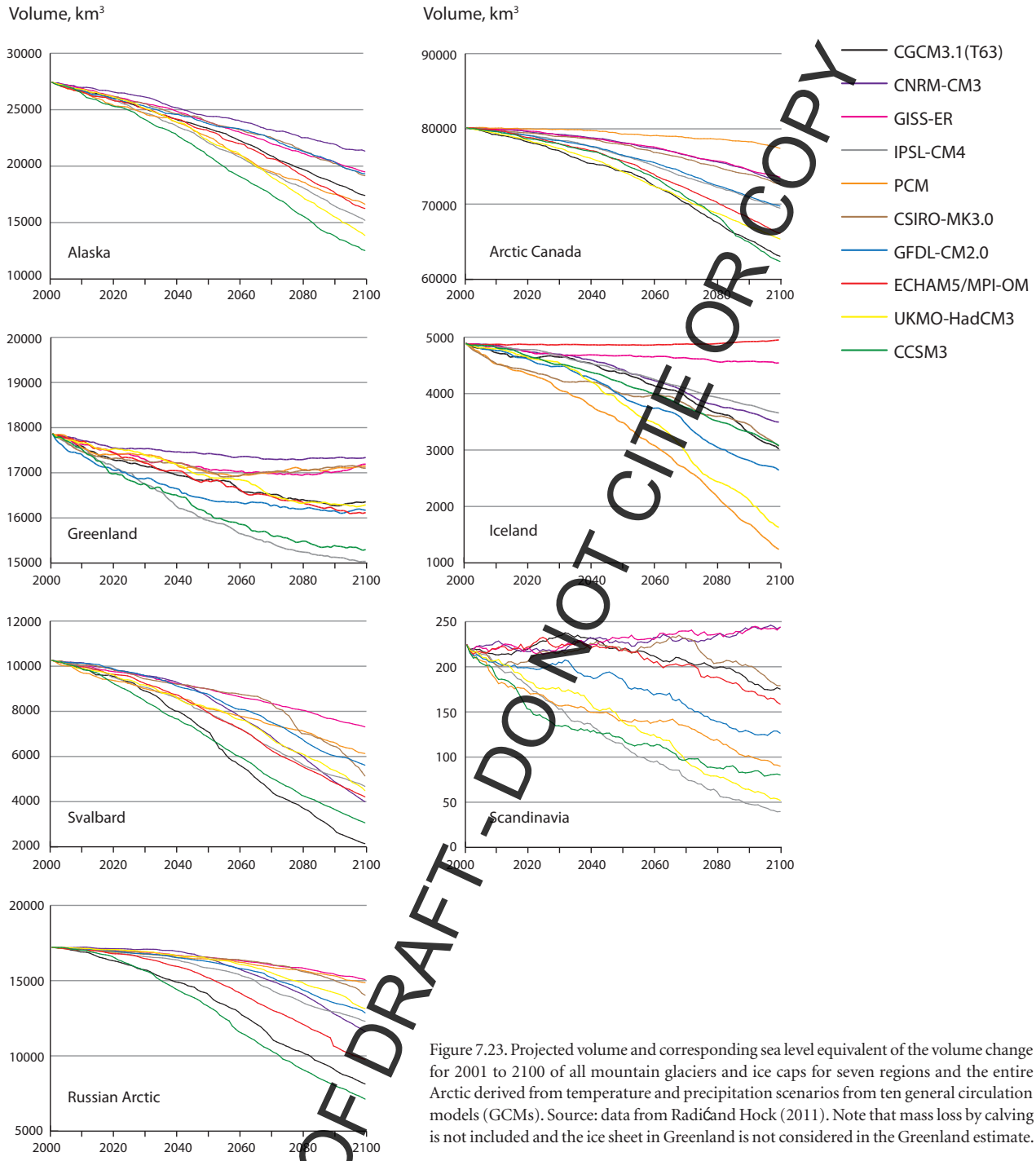


Figure 7.23. Projected volume and corresponding sea level equivalent of the volume change for 2001 to 2100 of all mountain glaciers and ice caps for seven regions and the entire Arctic derived from temperature and precipitation scenarios from ten general circulation models (GCMs). Source: data from Radić and Hock (2011). Note that mass loss by calving is not included and the ice sheet in Greenland is not considered in the Greenland estimate.

7.6.3. Modeling ice dynamics and ice extent

Glacier size and geometry will change as a glacier responds to climate change. Hence, projections of glacier mass balance on timescales longer than a few decades need to take into account feedbacks between the mass balance and the changing glacier surface elevation.

Ice flow models have been used to simulate area and thickness changes of individual glaciers (e.g., Schneeberger et al., 2003; Aðalgeirsdóttir et al., 2006). However, the detailed input data required to run such models are often not available. As a result, the effects of area changes are either simply neglected (e.g., Oerlemans et al., 2005; Zhang et al., 2007b) or approximated by scaling methods based on a functional relationship between glacier volume and glacier area (e.g., Radić and Hock, 2006). The total glacier volume is adjusted annually according to the volume changes computed from a climate-driven mass balance model. Annual area changes are then computed by inverting the volume-area relationship. Based on modeling of eleven glaciers, Schneeberger et al. (2003) showed that volume loss is overestimated by about 20% in a 100-year simulation if changes in glacier geometry are not included in the projections.

Detailed ice dynamic modeling is currently not possible on a regional scale due to a lack of the required data to run models for the vast majority of glaciers. Raper et al. (2000) developed a geometrical approach, in which the width, thickness, and length of a glacier are reduced as its volume and area decline. When applied statistically to the global population of glaciers and

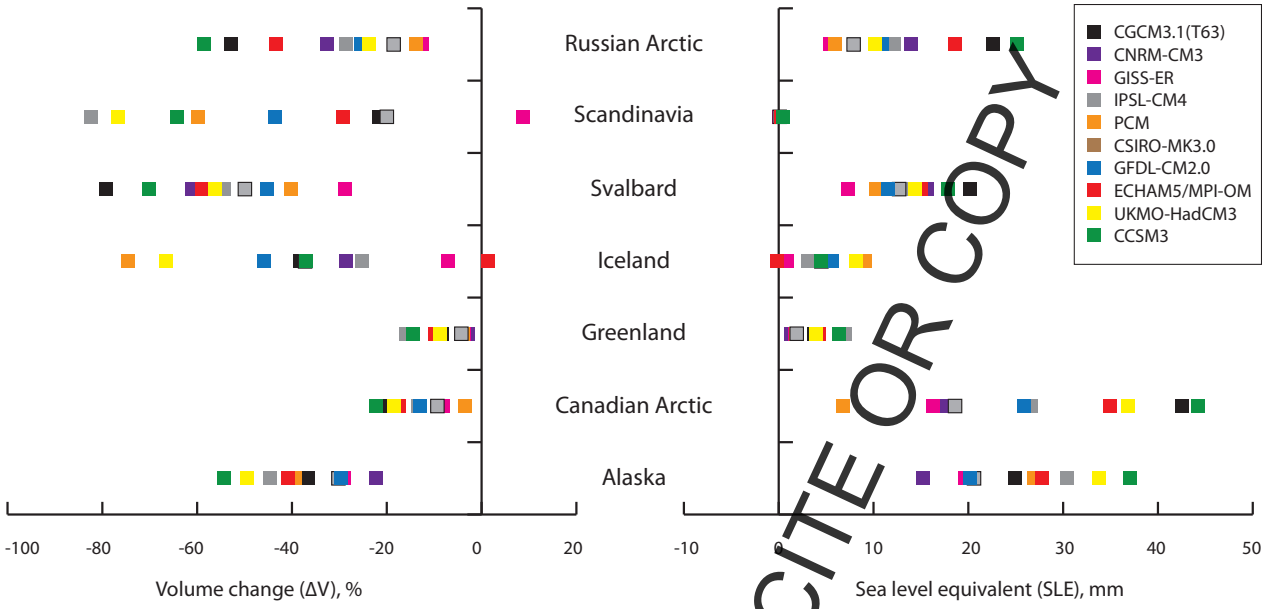


Figure 7.24. Projections of volume change and sea level equivalent of the volume change by 2100 for seven regions that include all Arctic glaciers. Projections are based on temperature and precipitation modeled by ten general circulation models (GCMs). Source: Radić and Hock (2011).

individually to ice caps, this approach shows that the reduction in glacier area strongly reduces ablation during the 21st century (Raper and Braithwaite, 2006). The reduction is about 45% under the IPCC A1B emissions scenario using output from two different GCMs.

A large number of glaciers in the Arctic terminate in the ocean and, therefore, lose mass through iceberg calving in addition to surface melting. Studies on a few Arctic marine-terminating ice caps (Dowdeswell et al., 2002, 2008; Burgess et al., 2005; Mair et al., 2009) indicate that calving may account for roughly 30% to 40% of total mass loss. However, model-based projections on regional scales neglect this effect. The current state-of-the-art in modeling does not allow meaningful projections of calving losses. However, calving rates are likely to increase as surface melting increases (Oerlemans et al., 2005).

7.7. Impacts of changes in mountain glaciers and ice caps

- The recent acceleration of mass loss from mountain glaciers and ice caps in the Arctic is such that, together with the Greenland Ice Sheet, they are likely to account for more than 60% of the current glacier wastage contribution to global sea level rise.
- Glacier wastage is likely to increase inputs of freshwater, sediment, and some nutrients to Arctic coastal waters, with potential impacts on water mass circulation, biological productivity, and ecosystem structure in affected fjord and nearshore marine environments.
- Retreat of tidewater glaciers onto land may be removing important feeding habitats that exist for seabirds and marine mammals in regions of upwelling in front of tidewater glacier termini and reducing the occurrence of grounded bergy bits that are used by seals for resting.
- Icebergs are a hazard to shipping and offshore platforms in the Arctic. The number and size distribution of bergs produced may change as tidewater glaciers retreat (e.g., more large bergs if floating ice tongues and ice shelves break up; more smaller bergs if calving from fast-flowing outlets increases). The way they circulate in Arctic waters is likely to change if sea-ice cover continues to decrease and surface water temperatures rise as a result. Bergs may become more mobile in summer as the restraining effect of sea ice declines, but disintegrate more rapidly in warmer surface waters.
- As climate warms, glacier runoff initially tends to increase due to higher melt rates, but ultimately declines as glacier area shrinks. In most regions of the Arctic, the phase of declining runoff does not yet seem to have begun, but it may begin soon in the Russian Arctic mountains.
- Changes in glacier runoff will ultimately impact the viability of hydroelectric power operations in the Arctic. They will also be associated with changes in stream temperature, sediment load, and nutrient content that will initiate changes in the ecology of downstream river and lake environments.
- As rates of mass loss increase, 'legacy pollutants' stored in firn and glacier ice will be released back into the environment.
- As glaciers retreat, there are likely to be changes in the magnitude and frequency of a range of geomorphological hazards, including mass movements from deglaciated valley walls and outburst floods from ice-marginal and proglacial (moraine-dammed) lakes.

7.7.1. Impacts on sea level

In the most recent decade for which mass balance data were available for all regions (1996 to 2006), Dyurgerov and Meier (2005) reported that Arctic glaciers collectively lost mass at a rate of 165 Gt/y (0.45 mm/y sea level equivalent; SLE). The dataset compiled by Cogley (2009a) yields a loss rate of 180 Gt/y (0.5 mm/y SLE) for the decade 1995 to 2005, and a loss rate of as much as 282 Gt/y (0.75 mm/y SLE) for the

2000 to 2005 pentad (see Section 7.9 for a more extended discussion). For comparison, Rignot et al. (2008) calculated a mass loss from Greenland of 267 Gt/y (0.71 mm/y SLE) for 2007. These results suggest that, together, Arctic glaciers and the Greenland Ice Sheet may account for more than 60% of the current glacier wastage contribution to global sea level rise (2.1 mm/y; Cazenave et al., 2009).

The acceleration of mass loss from Arctic glaciers and ice caps for the period 1985 to 2003 calculated using data from Dyurgerov and Meier (2005) is -7.3 km^3/y^2 ice equivalent, or 6.5 Gt/y^2 (0.02 mm/y^2 SLE) (Figure 7.25). Note that at this rate of acceleration the estimated reservoir of mountain glacier and ice cap sea level equivalent of 0.41 m (Radić and Hock, 2010) is being drawn down by only about 0.2% per year. While the current trend cannot be reliably extrapolated any significant distance into the future, depleting the reservoir is unlikely to result in a decline in annual total balance for at least the next several decades.

Analysis of the Cogley (2009a) dataset yields an estimate of mass losses from the glaciers surrounding the Greenland Ice Sheet of about 35 to 45 Gt/y for the 2000 to 2005 pentad. Some methods used to assess the mass balance of the Greenland Ice Sheet, such as most GRACE gravimetric observations, account for these losses, while other methods, such as satellite altimetric and mass flux observations, do not. It is therefore important, when combining mass loss estimates from this chapter with those in Chapter 8 on the Greenland Ice Sheet, to identify which glaciers are included in each particular method to avoid double counting.

7.7.2. Impacts on the marine environment

7.7.2.1. Freshwater and associated chemical fluxes

Glacier runoff is projected to increase in many parts of the Arctic for a period of decades or longer before declining (see also Section 7.7.3). This increased freshwater flux will influence conditions in coastal and marine waters. In fjords, freshwater discharge in summer produces a stable surface layer with low salinity (e.g., Svendsen et al., 2002), an effect that would be intensified with accelerated glacier melt. In the Gulf of Alaska, the majority (78%) of freshwater discharge occurs via a large number of smaller rivers and streams rather than a small number of point sources (Neal et al., 2010), producing baroclinic transport along the coast that is modified by wind stress (Royer, 1982). This along-coast transport is a major source of freshwater to the Bering Sea shelf and the Arctic Ocean (Weingartner et al., 2005). Runoff from glaciers and icefields contributes about half of the freshwater discharge to the Gulf of Alaska, with 10% currently from wastage (Neal et al., 2010). Because the magnitude of baroclinic transport varies with freshwater flux, future changes in glacier runoff volume and timing will have significant implications for the strength of the Alaska Coastal Current, associated marine ecosystems, and economically valuable fisheries.

In addition to changes in freshwater flux, changes in material export may influence coastal and marine ecosystems. Hood and Scott (2008) monitored three streams in southeastern Alaska and found that decreasing glacier cover was associated

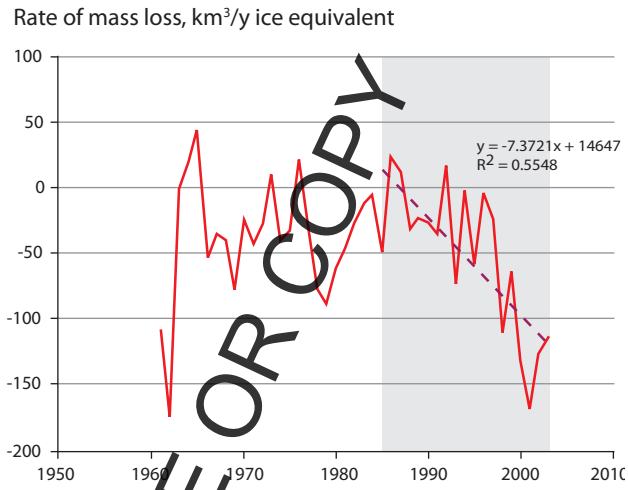


Figure 7.25. Rate of mass loss from Arctic glaciers and ice caps only, derived from Dyurgerov and Meier, (2005). Regression line fitted to the large square data points shows the acceleration of mass loss from 1985 to 2003. Source: AMAP.

with decreased phosphorus export and increased export of nitrogen and dissolved organic matter, although with a decrease in the more labile forms of organic matter. Nitrogen export increases in the early stages of vegetation succession on deglaciated forelands due to the establishment of nitrogen-fixing species such as alder, but is likely to decrease with the later establishment of other species such as conifers.

Increased glacier melt may be associated with increased sediment loads, which would increase turbidity in coastal waters. Increased turbidity would reduce the penetration of solar radiation, including harmful UV-B radiation, with implications for primary production and ecosystem functioning (Hanelt et al., 2001; Erga et al., 2005).

7.7.2.2. Seabird and marine mammal habitat

Tidewater glacier fronts are known to be 'hot spots' for seabirds and marine mammals in many Arctic areas. In Svalbard, a strong preference for these habitats has been identified for ringed seals (*Phoca hispida*) and beluga (*Delphinapterus leucas*) during summer and autumn (Lydersen et al., 2001; Freitas et al., 2008). Individuals of both species spend significant amounts of their time in proximity to tidewater glacier fronts, and their diving behavior in these areas suggests that they are feeding in these locations, which are known to be areas of high productivity. Enhanced production in these areas is probably due to freshwater outflows that drive upwelling of deep water close to glacier fronts, bringing nutrients to the surface and stimulating phytoplankton blooms and zooplankton growth. In addition, the cold-water outflows from the glaciers may stun or kill invertebrates, thus attracting predators such as polar cod (*Boreogadus saida*) and capelin (*Mallotus villosus*) that are an important part of the diet of beluga and ringed seal (Dahl et al., 2000; Labansen et al., 2007), as well as a variety of seabirds and other marine mammals. In addition, ringed seals, bearded seals (*Erignathus barbatus*), and other pinnipeds use bergy bits as resting platforms. Retreat of tidewater glacier fronts onto land would decrease the availability of this preferred habitat at a time when decreases in late summer sea-ice extent are making the other preferred summer feeding habitats of Arctic

pinnipeds less available. In addition, changes during glacier retreat in the lability of organic matter and the amount and types of nutrients that are delivered from land to the marine environment in meltwater runoff may reduce the productivity of these environments and their attractiveness to marine mammals and seabirds.

7.7.2.3. Iceberg production and drift

In some Arctic and sub-Arctic offshore areas, icebergs present a significant danger for marine operations including shipping and commercial fisheries, geological exploration surveys, and production of hydrocarbons from sea surface structures such as oil rigs (Figure 7.26). The Grand Banks of Newfoundland, the Labrador Sea, offshore of West Greenland, Davis Strait, and the Barents Sea are typical examples of such areas. Damage to or even destruction of a stationary platform or its communication lines by an iceberg will result in enormous economic losses and possibly significant environmental damage. Both iceberg plan form and keel depth are important quantities to measure in this context (Woodworth-Lynas et al., 1991). Non-stationary platforms and other floating structures (e.g., floating production, storage, and offloading systems), capable of leaving their operation localities, will limit economic losses owing to interruption of hydrocarbon production for the period of shutdown due to iceberg threat.

The general direction of the drift of icebergs near the Labrador coast and the Grand Banks of Newfoundland coincides with the direction of a strong Labrador Current. Iceberg trajectories and the most dangerous drift directions are known. Over the past few decades, Canada has gained significant experience in minimizing iceberg threats and has introduced a complex system of operational practices known as the Iceberg Management System.

Currents in the Barents Sea carry icebergs from the north to the southwest (Johannessen et al., 1999; Dmitriev and Nesterov, 2007). Some may enter the central and southern regions of the area, threatening navigation and operations in the *Shtokman* gas-condensate field. An analysis of atmospheric conditions at the time of the southernmost spreading of icebergs (e.g., 1881, 1929, 1989, and 2003) has revealed that all cases were characterized by prolonged periods (three to five months) of winds from the north (Buzin et al., 2008). Under such conditions, large numbers of icebergs concentrated near Spitsbergen and Franz Josef Land (their source) are exported southward. In some years (1989, 2003), icebergs reached the *Shtokman* gas-condensate field area and in particularly extreme cases (1881, 1929) reached the coast of northern Norway and the Kola Peninsula. Sightings of an anomalously large concentration of icebergs and bergy bits (around 100 pieces) in the *Shtokman* gas-condensate field area in May 2003 (Naumov et al., 2003) significantly influenced the design concept for the production complex. The boundary of iceberg drift in the Barents Sea has been displaced to the south over the past 60 years (Zubakin et al., 2006; Buzin et al., 2008), probably related to climate warming and a corresponding increase in the production of icebergs (Figure 7.27). The tendency for the ice edge in the Barents Sea to be displaced toward the north in summer during the past 10 to 15 years (Divine and Dick, 2006) may result in earlier and easier release of icebergs stuck in the straits of Franz Josef Land. Consequently, a larger number of icebergs may be in free drift and represent a threat to shipping and platforms. However, the life expectancy of the icebergs also may be reduced in a warmer climate (Weeks and Campbell, 1973).

The iceberg discharge from the Greenland Ice Sheet is known to have increased by more than 40% from 1995 to 2005 (Rignot and Kanagaratnam, 2006). This has not, however, increased iceberg numbers on the Grand Banks or in the Labrador Sea. The number of icebergs reaching the Grand Banks has actually decreased, probably owing to increased surface water temperatures in the region and decreased ice cover (which protects icebergs from the dynamic impact of ocean waves) (McClintock et al., 2007). However, potential iceberg damage to sea-floor engineering structures is a function of iceberg keel depth as well as iceberg number. Icebergs with keel depths greater than 500 m have been observed in the Scoresby Sund fjord system of East Greenland, but few other systematic measurements of the keel depth of Arctic icebergs exist (Dowdeswell et al., 1992). The occurrence of recent plow marks on the sea floor, produced where drifting icebergs go aground, is evidence of this natural hazard, although many plow marks on Arctic continental shelves are not produced by modern icebergs (Dowdeswell et al., 1993; Syvitski et al., 2001).

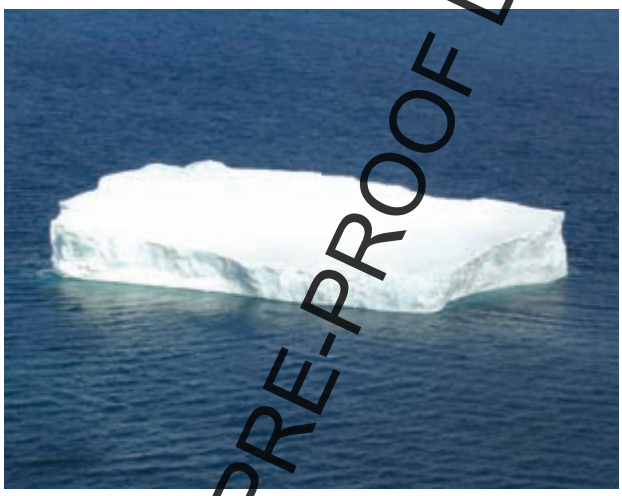


Figure 7.26. Large iceberg (100 m length, 30 m width, and 15 m height above sea surface) drifting in the Barents Sea to the southwest of Franz Josef Land on 30 May 2009. Source: Andrey Masanov, Arctic and Antarctic Research Institute, Russian Federation.

7.7.3. Impacts on water resources

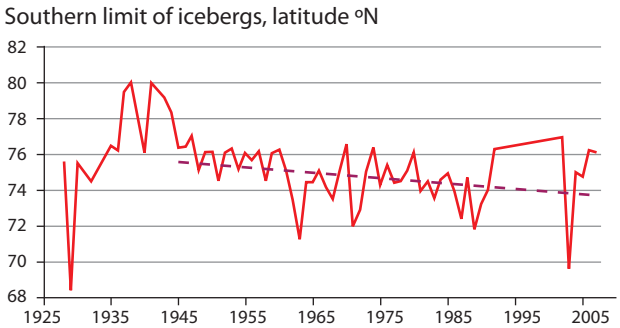


Figure 7.27. Multiyear variability of the southern limit of icebergs in the Barents Sea for the period 1928 to 2008. Source: Buzin et al. (2008).

7.7.3.1. Glacier runoff

Streamflow in glacier-fed streams tends to peak later in the summer and to continue to exhibit diurnal fluctuations later in the melt season than streamflow in nival (snowmelt-dominated) streams (Church, 1974). However, the contrast between nival and proglacial regimes can be obscured by the weather patterns dominating the particular melt season, particularly when melt begins later than normal (Marsh and Woo, 1981). Research in temperate areas and in the Arctic indicates that detectable glacier melt contributions to streamflow occur in catchments with more than about 5% glacier cover, while contributions to summer flow are detectable even in catchments with as little as 2% glacier cover (e.g., Marsh and Woo, 1981; Stahl and Moore, 2006). Based on these criteria, glaciers are an important water resource throughout much of the Arctic.

Over annual or longer timescales, the influence of glaciers on streamflow can be assessed through the catchment water balance, which can be expressed as follows:

[1] $Q = P - E - f_g B + GW_{net} - \Delta S$

where Q is streamflow, P is precipitation, E is evapotranspiration (including sublimation), f_g is the fractional glacier area within the catchment, B is the net mass balance for the glaciers, GW_{net} is the net flux of groundwater across the catchment boundaries, and ΔS represents changes in storage of seasonal snow, subsurface water, and surface water. All terms in [1] are typically expressed as a depth of water averaged over the catchment area. Negative net mass balance (glacier loss) augments streamflow relative to a basin lacking glacier cover. This contribution to streamflow ($f_g B$) is commonly referred to as 'wastage runoff'.

Table 7.3 highlights the contributions of glacier wastage (negative mass balance) to mean annual streamflow over recent decades, based on published water balance calculations and hydrological modeling. The significance of glacier melt contributions to streamflow tends to be greater than average in years with low precipitation and warm summers. At De Geerdalen, Svalbard, for example, glacier wastage contributed up to 20% of annual runoff, roughly double the average, in a year when annual precipitation was 40% below average (Killingtveit et al., 2003).

Assuming no further climate change, glaciers currently experiencing negative mass balance would recede until they achieved a hypsometry for which $B = 0$, as long as the ELA is below the top of the glacier. If the ELA lies above the maximum elevation of the glacier, the glacier would ultimately disappear. Under this assumption of constant climate, average annual runoff would ultimately decrease by an amount equal to the current contribution due to glacier wastage. Decreases in runoff would be particularly severe in years with low precipitation. Flowers et al. (2005) simulated the future evolution of Vatnajökull, Iceland, assuming a continuation of the 1961 to 1990 climatology, and found that glacier runoff would decrease by about 20% after 200 years.

Climatic warming will result in a longer melt period and more intense melt, particularly as firn cover is depleted, exposing lower-albedo glacier ice to solar radiation. As a consequence, glacier runoff will increase with the onset of a warming trend. However, increased melt is accompanied by glacier thinning and retreat, and at some point the decrease in glacier area will cause glacier runoff to decrease. While this conceptual model is widely recognized (e.g., Jansson et al., 2005; Moore et al., 2009), the timescales associated with the transition from increasing to decreasing flow are uncertain. Available studies focused on southern Alaska, northeastern British Columbia, and southwestern Yukon have reported positive trends in streamflow for glacier-fed rivers, suggesting that the glaciers in those regions are still in the initial phase of their response to warming (Fleming and Clarke, 2003; Brabets and Walvoord, 2009).

Runoff from glaciers in the temperate and sub-Arctic regions of Russia was calculated from the USSR Glacier Inventory for the period 1966 to 1980 based on the assumption that glacier systems are in equilibrium with climate. The total glacier runoff over the period 1966 to 1980 was estimated to be no more than 3 Gt (Ananicheva and Krenke, 2007). To project future changes in glacier runoff in Russia, the morphology and regime of glacier systems were adjusted on the basis of predicted changes in ELA under a climate-warming scenario derived from the ECHAM4 GCM driven by the IPCC A2 emissions scenario. Despite a significant reduction of the glacierized area, the glacier runoff from southeastern Siberia and the Sredinny Range (Kamchatka) was predicted to increase through the 21st century owing to increased ablation. Glaciers in the Kronotsky Range (Kamchatka) were predicted to almost disappear, resulting in a sharp reduction in glacier-melt runoff (Ananicheva and Krenke, 2007).

Transient responses of glacier geometry and runoff to climate warming have been simulated for Hofsjökull and Vatnajökull, Iceland (Flowers et al., 2005; Aðalgeirsdóttir et al., 2006), and the results are consistent with the conceptual model presented above. Figure 7.28 illustrates that, as the rate of warming increases, the interval with maximum glacier runoff

Table 7.3. Runoff contributions associated with glacier wastage.

Catchment	Location	Period	Catchment area, km²	Glacier coverage, %	Wastage contribution to streamflow, %	Q, kg/m²	Source
Zackenberg	Northeast Greenland	1997 – 2005	512	20	63	428	Mernild et al., 2008a
Mittivakkat	Ammassalik Island, Greenland	1993 – 2004	14.2	78	23	1970	Mernild et al., 2008b
Bayelva	Svalbard	1990 – 2001	30.9	55	23	1050	Killingtveit et al., 2003
De Geerdalen	Svalbard	1990 – 2001	34.4	17	9	539	Killingtveit et al., 2003
Endalen	Svalbard	1990 – 2001	28.8	20	19	545	Killingtveit et al., 2003

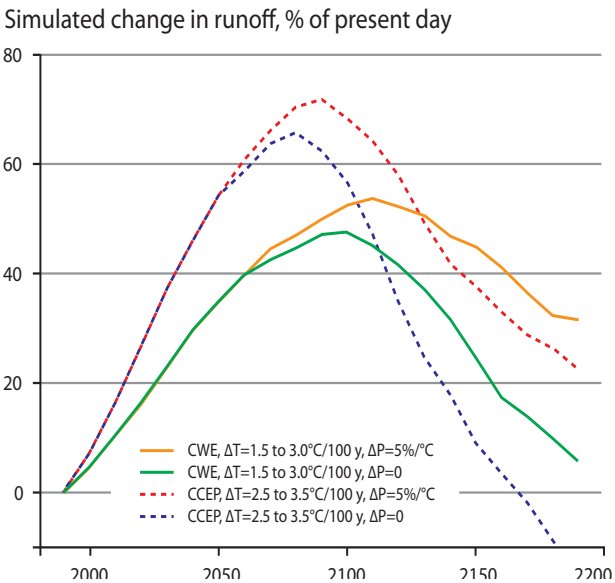


Figure 7.28. Simulated variations in glacier runoff from Hofsjökull, Iceland, under four climate scenarios. Source: Aðalgeirsdóttir et al. (2006).

for Hofsjökull moves forward in time and the increase in runoff is greater. For simulations involving a warming-related increase in precipitation, the effect is to increase the magnitude of runoff increase and delay the timing of peak runoff. Flowers et al. (2005) also reported that changes in the surface topography of the ice cap would cause shifts in the hydrological divides, effectively redistributing outflow among the outlet glaciers.

Climate warming and glacier response will modify seasonal patterns of runoff in addition to changing annual totals. Jónsdóttir (2008) simulated runoff in Iceland for climate projections based on the IPCC A2 and B2 emissions scenarios and projections of glacier geometry using a dynamic glacier model (Aðalgeirsdóttir et al., 2006; Jóhannesson et al., 2006). For the period 2071 to 2100, glacier-melt contributions to streamflow are projected to increase in all seasons, while runoff from unglacierized areas increases from October to April and decreases from May to September, jointly producing a significant shift in the seasonal pattern of runoff in partially glacier-fed catchments.

The highest peak flows in proglacial streams are often associated with outburst floods (Ng et al., 2007), in which water stored on, under, or within the glacier is released suddenly, for example, by the rapid growth of a meltwater channel (e.g., Mernild et al., 2008a). There is some evidence that subglacially derived outburst events are more extreme in warmer summers when there is greater meltwater generation (Skidmore and Sharp, 1999), suggesting that flood risk in glacier-fed Arctic catchments may increase as a result of climate warming. When outburst floods are derived from ice-dammed marginal lakes, thinning of the ice dams due to increased melting may result in floods that occur earlier in the melt season and are smaller and more frequent than they are at present. Lakes will fill sooner and more rapidly than in the past and the accumulated lake water may breach the ice dam more easily than before. Earlier onset of lake drainage may allow multiple filling/drainage cycles during a melt season, especially because melt seasons will be longer in a warmer climate.

7.7.3.2. Water quality

Climate warming will influence proglacial water quality as well as streamflow. Increasing meltwater runoff in the initial phases of warming should tend to moderate diurnal increases in stream temperature. However, as glaciers retreat, the length of stream exposed to solar radiation and other heat inputs will increase, potentially counteracting the effect of increased flow. In addition, the emergence of proglacial lakes can increase warming. Milner et al. (2008) found that maximum stream temperatures at Wolf Point Creek, which drains into Glacier Bay, Alaska, were 2 °C in 1977, roughly 30 years after the creek emerged due to glacier retreat. From 1996 onward, maximum temperatures have consistently exceeded 15 °C, at least in part due to the increase in size of the upstream lake, which initially formed in the 1970s. Once glaciers reach the stage of declining flow, there should be a trend to increasing stream temperature. At that point, the main factor mitigating this increase would be development of riparian forests where they do not currently exist, which would increase stream shading and enhance bank stability, resulting in a narrower, deeper stream. Both effects would tend to reduce stream warming, but would be relevant only in sub-Arctic and temperate areas such as the Gulf of Alaska.

Focusing on temperate regions, Moore et al. (2009) concluded that the onset of negative mass balance and glacier retreat would be likely to result in an initial increase in suspended sediment concentrations in proglacial streams, followed by a longer-term decrease. This long-term projection is consistent with studies that show that glacier-fed streams in the Arctic typically yield orders of magnitude more sediment than those without glacier contributions (e.g., Hasholt and Mernild, 2008). However, temporal changes in sediment concentration could vary substantially among basins due to site-specific factors, particularly the emergence of proglacial lakes, which trap sediment and thus reduce downstream transport (Hasholt et al., 2008). Future patterns of suspended sediment concentration would also be complicated in cases where glaciers shift from being cold-based or polythermal to temperate, as well as by thawing of permafrost, which could influence the erosion of stream channel beds and banks (McNamara and Kane, 2009).

Glacier retreat may be accompanied by changes in streamwater chemistry. Glacial forefields are subject to different weathering and nutrient cycling processes than those that dominate in subglacial environments, particularly as soils and vegetation develop (Anderson, 2007; Hood and Scott, 2008). Available studies, not all from Arctic areas, suggest that concentrations of nutrients and dissolved organic carbon should increase as glaciers retreat, although the forms of some nutrients and organic matter might change, possibly influencing their availability to aquatic organisms (Lafrenière and Sharp, 2005; Filippelli et al., 2006; Hood and Scott, 2008). In southeastern Alaska, decreasing basin glacier cover was found to be associated with higher concentrations of soluble reactive phosphorus and higher proportions of bioavailable (labile) dissolved organic carbon (Hood and Berner, 2009; Hood et al., 2009). Fluvial transport of dissolved organic carbon could be influenced by changes in suspended sediment concentration, because high turbidity reduces photodegradation of dissolved organic carbon in streamwater.

7.7.3.3. Water resources and hydroelectric power generation

Most studies suggest that Arctic glaciers are currently in the initial phase of their response to warming, in which runoff increases, and that glacier runoff should continue to increase over the next few decades or so, providing an increase in water availability. However, some Russian results indicate that some areas (especially those where the glaciers are predominantly small) are likely to experience flow decreases over the next few decades as a result of glacier retreat (Ananicheva and Krenke, 2007). These contrasting behaviors could either reflect fundamental differences among glaciers in the timescale of response to warming, or suggest that glaciers are currently at different stages of their response. Further site-specific assessments are clearly warranted.

Glacier runoff is significant for hydroelectric power generation in the upper Yukon basin (at Whitehorse, Yukon, Canada), Iceland, Greenland, and in many river systems in Norway. A trend to increasing flows would also tend to increase the hydroelectric power resource. In Iceland, Jónsdóttir (2008) projected a substantial increase in hydroelectric power potential for the period 2071 to 2100, but a substantial decrease over the following century as the glaciers disappeared under the IPCC A2 and B2 emissions scenarios employed.

Reservoir operations could be influenced not only by the effects of changes in streamflow, but also by changes in sediment transport and stream temperature. Increased sediment transport would reduce the operational lifetime of a reservoir by infilling, while suspended sediment concentration and inflow temperature together would influence mixing processes and temperature profiles, in turn influencing habitat conditions for aquatic organisms as well as the temperature of water downstream.

7.7.3.4. Aquatic ecosystems

Glacier-fed rivers typically support a distinctive flora and fauna determined by physicochemical conditions, including the dominant influences of water temperature and channel stability (Milner and Petts, 1994; Milner et al., 2001). While bedload and suspended sediment transport typically peak during maximum glacial melt in summer (Østrem, 1975b), lower water levels during spring and autumn result in greater water clarity and channel stability, creating 'windows of opportunity' when algal and macroinvertebrate productivity is higher. Some of the largest runs of salmon in south-central Alaska occur in glacially influenced systems (Deravy and Milner, 2000) as glacial runoff enhances summer flow relative to rainfall- or snowmelt-dominated catchments, facilitating the migration of salmon from the ocean to their spawning grounds in tributary streams via the main channel corridor.

Since increases in glacial runoff will enhance summer flow and increase the duration of meltwater runoff, they will probably also facilitate the migration of adult salmon to their spawning grounds. Increased glacial runoff will also increase the amount of available spawning habitat and would potentially enhance winter baseflow through recharge of groundwater during the summer (Smith et al., 2001; Brown et al., 2003). The survival of young salmonids during the winter is significantly correlated with the amount of winter discharge or winter groundwater inputs (Fleming, 2005). Any changes in water source contributions that enhance winter flow should be beneficial to salmonid populations and mitigate the effect of overwintering mortality on salmonids.

In the phase of declining glacier runoff, bedload and suspended sediment transport will potentially decrease in the long term as glacial and paraglacial erosion (erosion by non-glacial processes of previously deposited glacial and fluvioglacial sediments) decrease and the stream's transport capacity decreases (Fleming and Clarke, 2005). In addition, newly formed proglacial lakes trap sediment, reducing sediment loads downstream. Fleming (2005) emphasized the potentially favorable effects of glaciers and lakes in combination on salmonid habitat: glaciers maintain late summer flow while lakes reduce sediment transport. Reduced bedload transport could allow streams to become more stable, and thus narrower and deeper. The projected reduction in suspended sediment load with ongoing glacier retreat should allow greater light penetration into the water column and increase algal productivity, particularly during certain periods of the year. The effect of increased light would be augmented by an increase in nutrient concentrations as vegetation and soils develop following glacier retreat. However, the effects of decreased turbidity on ecosystem productivity need to be considered in the context of possible decreases in late summer streamflow and increasing late summer water temperature.

There is clear evidence of a strong relationship between catchment glacierization and macroinvertebrate taxon richness in glacially influenced streams. In a stream fed by a receding glacier in southeastern Alaska, Milner et al. (2008) found that the benthic macroinvertebrate community supported only five taxa when 70% of the catchment was glacierized but 24 taxa after the glacier had disappeared two decades later. Füreder (2007) found similar trends in alpine streams with differing degrees of glacierization. This evidence suggests that changes in glacial runoff and fluvial sediment transport due to glacier shrinkage may cause an increase in the diversity of the benthic macroinvertebrate community. An increase in algal productivity will potentially enhance macroinvertebrate abundance. With a change in benthic community structure, dominant species traits will reflect a reduction in glacial contribution to flow. Macroinvertebrate body size will become larger and life cycles longer, with the benthic community dominated by habitat generalists with less omnivory and flattened forms (Ilg and Castella, 2006; Füreder, 2007). The fauna would also be less cold-adapted with reduced mobility of the adults. Enhanced abundance and increased body size will favor food availability for fish. Although an overall increase in taxonomic richness and abundance in benthic macroinvertebrate communities is predicted with decreased glacial inputs, some cold-specialist benthic species will be lost, including some endemic species. Some species will not adapt rapidly enough to survive the anticipated rate of deglaciation in future decades (Hari et al., 2006).

The effect of shrinking glaciers on fish populations in glacially influenced rivers will depend upon whether the system is an alpine or a more lowland system. However, over time reduced glacial input into these rivers will reduce summer flow and may influence salmon migration upstream. Lower flow

will also reduce the amount of habitat available adjacent to the main glacier-fed channel, including side channels and side sloughs. Side channels and sloughs are sustained by meltwaters of the main channel (Richardson and Milner, 2005). For many river systems in northerly regions, winter is a critical period when overwintering mortality for salmonids can be high if they are unable to migrate to suitable refugia, such as side channels. Water temperature is a critical variable influencing fish distribution and each species has an optimum range for different life stages. Matulla et al. (2007) predicted that the ranges of alpine stream salmonids (e.g., brown trout *Salmo trutta* and grayling *Thymallus thymallus*) would contract upstream due to changes in water temperature regimes under climate change scenarios in which the cold water input from glacial runoff decreases. Projected changes in water temperature in cold rivers can also favor non-indigenous species that have wider thermal tolerance and can out-compete indigenous taxa at higher water temperature (Ficke et al., 2007). In the Rocky Mountains, cutthroat trout (*Oncorhynchus clarkii*) have been displaced by non-indigenous brook trout (*Salvelinus fontinalis*) and brown trout that are more efficient at securing food resources at higher temperatures (Hauer et al., 1997). However, in streams that are currently colder than is optimal for salmonid growth, some degree of warming might increase salmonid production.

Large-scale glacial recession creates new habitat that can potentially be colonized by organisms including fish. The creation of new stream habitat due to glacial recession has been extensive along the coasts of southeastern and south-central Alaska due both to climate change and changing local environmental conditions (Milner et al., 2008). Although many of the recently formed streams are relatively short, the creation of new salmonid habitat is significant at a time when human activity threatens salmonid stocks in other parts of the Pacific Northwest region of North America (Nehlsen et al., 1991). Pink salmon (*Oncorhynchus gorbuscha*) can establish substantial populations from a small colonizing population of spawners within a few generations, and other species (including non-anadromous fish such as sticklebacks) colonize rapidly (Milner et al., 2008). As migrant salmon die soon after spawning, rotting of their carcasses results in a significant input of nutrients and organic matter to affected catchments, and this can promote productivity in other parts of the ecosystem.

7.7.3.5. Contaminant release from melting glaciers

Many contaminants originating from lower latitudes are carried to the Arctic by long-range atmospheric transport and stored in the firn and ice of glaciers and ice caps (see also section 11.3). These contaminants include sulfate and nitrate, which have increased the acidity of snow and ice above pre-industrial levels (Koerner and Fisher, 1982), metals such as lead (Zheng et al., 2003), persistent organic pollutants such as PCBs (polychlorinated biphenyls), DDT and other organochlorine pesticides (Gregor and Gummer, 1989; Hermanson et al., 2005), as well as radionuclides associated with the Chernobyl reactor failure in 1986 and with nuclear testing dating back to 1961 to 1962 (Pinglot et al., 1999). Stored contaminants may be released back into the environment by glacier melt and the process can be accelerated if melt rates increase over time (Blais et al., 2001).

Contaminants stored in seasonal snow away from glaciers interact with soil and aquifer material prior to their discharge to receiving waterbodies such as lakes and streams. During these interactions, a proportion of the contaminants (especially persistent organic pollutants) released via melting will be retained through sorption onto organic matter and cation exchange on clay particles in the soil. In contrast, glacier meltwater has much less interaction with organic matter and clays in soils before leaving the glacier, leading to less retention and higher rates of contaminant export to lakes and rivers downstream, as has been found in the Canadian Rockies (Blais et al., 2001; Lafreniére et al., 2006).

All the contaminants stored in glaciers would eventually be discharged in meltwater runoff regardless of climate change, but there would be significant lag times associated with the flow of ice from the accumulation to ablation zones. Climate warming accelerates melting of snow, firn, and ice, potentially releasing stored contaminants to receiving waters in a more concentrated pulse. However, thus far there has been no research in the Arctic directly addressing this impact.

7.7.4. Glacier ecosystems

Distinctive microbial ecosystems are found on and under glaciers (Hodson et al., 2008). On the glacier surface, the supraglacial ecosystem exists in the snowpack, supraglacial streams, melt pools, and cryoconites (water-filled holes on glacier surfaces that contain dust and microbial communities). It comprises a diverse array of bacteria, algae, fungi, viruses, rotifers, and tardigrades. *In situ* rates of primary production and respiration on glacier ice surfaces in summer are comparable to those encountered in soils in warmer and more nutrient-rich regions. Measurements of net production suggest that glacier ice surfaces may be largely autotrophic systems (Anesio et al., 2009), unlike most lakes and rivers, which tend to be heterotrophic systems.

At the glacier bed, a subglacial ecosystem is found in basal ice, subglacial sediments, meltwaters, and volcanic subglacial lakes (Gaidos et al., 2004, 2009; Hodson et al., 2008). It includes a mixture of aerobic and anaerobic (chemoautotrophic and heterotrophic) bacteria and probably also fungi and viruses (D'Elia et al., 2008). These microorganisms have metabolisms based on the oxidation of organic carbon, sulfide, sulfur, or hydrogen, and they use oxygen, nitrate, sulfate, iron(III), manganese(IV), or carbon as terminal electron acceptors. They play a critical role in subglacial chemical weathering (Tranter et al., 2002, 2005) and in controlling the biogeochemistry, nutrient loading, and properties of dissolved organic matter in glacial runoff (Hodson et al., 2008; Gaidos et al., 2009). They may also be involved in the production of greenhouse gases (carbon dioxide and methane), with possible feedbacks to the climate system if these gases are stored in basal ice or subglacial sediments and released to the atmosphere during deglaciation (Sharp et al., 1999; Skidmore et al., 2000; Wadham et al., 2008; Boyd et al., 2010).

The impact of climate change on these ecosystems is likely to be complicated as there will be a number of competing influences (Hodson et al., 2008). In the long term, changes in global glacier area will limit their geographical extent, and

total deglaciation would be likely to eradicate these ecosystems completely. In the shorter term, increased surface melt would probably enhance supraglacial ecosystem productivity. As end-of-summer snowline elevations rise, the extent of supraglacial stream and meltwater habitats will increase at the expense of snowpack habitats. Higher ice melt rates may swamp these systems with water, continually redistributing them across the glacier surface. Increased melting of glacier ice may increase the rate of release of glacially entombed microorganisms into the supraglacial environment. This may increase the transfer of carbon, nutrients, and biota into downstream (subglacial and proglacial) aquatic environments. Increased water fluxes would be likely to increase the oxygenation of glacier beds in summer, favoring aerobic over anaerobic bacterial processes. They might also result in more rapid transit of meltwater through the subglacial ecosystem, especially as glaciers shrink, reducing the efficiency of utilization of supplies of nutrients and organic matter by subglacial microbial communities. Alternatively, increased delivery of carbon and nutrients could stimulate microbial activity, and create anoxic conditions at glacier beds in winter when surface water inputs cease.

Paradoxically, however, negative surface mass balances and glacier thinning have resulted in cooling and net freezing at the beds of some Arctic glaciers (Dowdeswell et al., 1995; Hodgkins et al., 1999; Glasser and Hambrey, 2001). Such changes would lead to isolation and cryostasis (the reversible cryopreservation of live organisms) of the subglacial ecosystem. Whether this might ultimately eliminate the subglacial ecosystem from individual glaciers or drive it deeper into unfrozen subglacial sediments is currently unknown. Elimination of the ecosystem would substantially alter the biogeochemistry of glacial runoff, with potential implications for downstream ecosystems (Hodson et al., 2004, 2005).

7.7.5. Geomorphological hazards

Glacier changes, particularly glacier retreat, can produce a range of hazards involving slope instability, increased proglacial fluvial sediment transport, and changes in water flow, particularly outburst floods. To augment the available research conducted in the Arctic, the review below draws heavily on work outside the Arctic. However, the general principles should apply throughout the Arctic region as broadly defined in this chapter.

7.7.5.1. Slope instability

Glacier down-wasting and retreat appear to be partly responsible for destabilizing adjacent terrain, leading to landslides, debris flows, rock avalanches and falls, and some catastrophic rock-slope failures in high mountains (Evans and Clague, 1994; Holm et al., 2004). Many marginally stable slopes that were buttressed by glacier ice during the Little Ice Age failed after they became deglaciated in the 20th century. A factor that has possibly contributed to such failures is steepening of rock slopes by cirque and valley glaciers during the Little Ice Age. These effects are most pronounced in mountain ranges with the largest ice cover, because it is there that ice losses in the 20th century have been greatest. An extreme example is Glacier Bay, which until the end of the 18th century was covered by glacier ice. Since then, Glacier Bay has become deglaciated, with the loss of over 1000 km^2 of ice in 200 years. The amount of ice lost is so great that the uplift of the land due to isostatic rebound is measurable (Larsen et al., 2005).

7.7.5.2. Outburst floods from moraine- and ice-dammed lakes

Lakes dammed by moraines and glaciers in high mountains have drained suddenly to produce floods orders of magnitude larger than normal nival or rainfall floods (Costa and Schuster, 1988; Clague and Evans, 1994). Lakes dammed by Neoglacial end and lateral moraines are susceptible to failure because they are steep-sided and consist of loose, poorly sorted sediment that in some cases is ice-rich (Clague and Evans, 2000). Irreversible rapid incision of a moraine dam may be caused by a large overflow triggered by an ice or snow avalanche or rockfall. Other failure mechanisms include earthquakes, slow melt of ice within the moraine, and slow removal of fine-grained sediment from the moraine by groundwater flow (piping). As climate warms, lakes impounded by glaciers may drain suddenly and unexpectedly following a long period of stability due to progressive wastage of the glacier dam and the formation of subglacial, supraglacial, or ice-marginal channels. Drainage may be initiated by hydrostatic lifting of that part of the base of the glacier facing the lake. Lakes may also form during glacier surges; they drain, often catastrophically, soon after they form.

Most outburst floods display an exponential increase in discharge, followed by a gradual or abrupt decrease to background levels as the water supply is exhausted (Clarke, 2003; Ng et al., 2007). Peak discharges are controlled by lake volume, dam height and width, the material properties of the dam, failure mechanism, and downstream topography and sediment availability. Floods from glacier-dammed lakes tend to have lower peak discharges than those from moraine-dammed lakes of similar size because enlargement of tunnels within ice is a slower process than overtopping and incision of sediment dams.

Floods resulting from failures of natural dams may transform into debris flows as they travel down steep valleys. Such flows can only form and be sustained on slopes greater than 10 to 15 degrees and only where there is an abundant supply of sediment in the valley below the dam. Entrainment of sediment and woody plant debris by floodwaters may cause peak discharge to increase down valley, which has important implications for hazard appraisal.

Outburst floods from lakes dammed by moraines and glaciers commonly alter river floodplains tens of kilometres from the flood source. The floodwaters erode, transport, and deposit huge amounts of sediment. They commonly broaden floodplains, destroy pre-flood channels, and create a new multi-channel, braided to anastomosing (channels that repeatedly branch and reconnect) plan form. The changes can persist for decades after the flood, although rivers quickly reestablish their pre-flood grades by incising the flood deposits.

Climate is an important determinant of the stability of moraine and glacier dams (O'Connor and Costa, 1993; Evans and Clague, 1994). Most moraine-dammed lakes formed during the past century as glaciers retreated from bulky end moraines constructed during the Little Ice Age. The lake dams soon began to fail as climate warmed. With continued warming and glacier retreat, the supply of moraine-dammed lakes that

are susceptible to failure will be exhausted, and the threat they pose will diminish. Glacier-dammed lakes have typically gone through a period of cyclic or sporadic outburst activity, lasting up to several decades, since climate began to warm in the late 19th century. The outburst floods from any one lake ended when the glacier dam weakened to the point that it could no longer trap water behind it. However, with continued glacier retreat, the locus of outburst activity may, in some cases, shift up-glacier to sites where new lakes develop in areas that are becoming deglaciated. Landslide damming is less clearly linked to climate.

7.7.5.3. Erosion and sedimentation

Fluctuations of glacier margins on timescales of decades and centuries can remobilize glacial sediment. During glacier advance, initial incision due to increased competence of meltwater streams is quickly followed by deposition as sediment supply increases (Maizels, 1979). Sediment stored within and beneath glaciers is delivered at an increasing rate to fluvial systems as glaciers advance (Karlén, 1976; Maizels, 1979; Leonard, 1986, 1997; Karlén and Matthews, 1992; Lamoureux, 2000). Similarly, subglacial erosion increases during glacier advance, and meltwater may carry more sediment into river valleys than at times when glaciers are more restricted. Sediment delivery to streams in the Coast Mountains of British Columbia, for example, increased during the Little Ice Age, and the streams responded by filling in their channels and braiding over distances up to tens of kilometres down valley from glaciers (Church, 1983; Gottesfeld and Johnson-Gottesfeld, 1990). Glacier retreat typically exposes large areas of unstable, unvegetated sediment that is easily entrained by meltwater (Church, 1983; Desloges and Church, 1987; Gottesfeld and Johnson-Gottesfeld, 1990; Brooks, 1994; Ashmore and Church 2001; Clague et al., 2003).

7.7.5.4. Glacier-volcano interactions and related hazards

Glaciers are found in regions of active volcanism in Alaska, Iceland, and Kamchatka, where high seismic and volcanic activity can generate significant glacier-related hazards. Consequences of volcanic activity include glacier burial by ash falls, melt-induced formation of craters and lakes on glaciers, significant mass loss by subglacial melting during subglacial eruptions, and accelerated glacier flow. Several volcanogenic *jökulhlaups* (outburst floods from glaciers) have been documented in Iceland (Russell et al., 2006).

Lahars (fluidized volcanic debris flows produced when hot pyroclastic material falls onto glaciers) can be extremely destructive owing to their high density and low viscosity while in motion, and their tendency to over-consolidate (stiffen) upon halting. Lahars are especially destructive because their mode of origin allows large volumes of material to be activated quickly, and because the topographic relief of the volcano provides abundant potential energy. In Kamchatka, lahars have caused significant damage to infrastructure and the environment. For example, the 2001 eruption of Shiveluch Volcano produced lahars which traveled about 20 km and damaged the road to Ust-Kamchatsk. The 2005 eruption of Kluchevskaya Volcano created lahars that carried blocks 3 to 4 m in size a distance of over 30 km.

It is unclear from the available literature how glacier-volcano interactions and the associated hazards will be influenced by future changes in climate and resulting changes in glaciers and ice caps. However, there is some evidence that volcanic activity may increase as ice thins (Tuffen, 2010).

7.7.5.5. Glacier surges

Surging glaciers are found in Alaska, the Yukon Territory, Arctic Canada, Greenland, Iceland, Svalbard, and Novaya Zemlya. Those in Novaya Zemlya tend to terminate in water (Grant et al., 2009), and surges of these glaciers are a potential danger to local populations and infrastructure. Some surging glaciers in Alaska, such as the Black Rapids Glacier, are a potential hazard to the Trans-Alaska (oil) Pipeline (Heinrichs et al., 1995). Surge behavior has been documented for 26 glaciers in Iceland (Björnsson et al., 2003). Surges of these glaciers are associated with an increase in proglacial stream turbidity, but not discharge. The geographical variability of surge behavior, coupled with the incomplete understanding of glacier surge mechanisms, makes it difficult to predict whether climate change will alter the frequency of surging or the impact of the phenomenon on human society.

7.7.6. Glacier-related tourism

More than 1.5 million tourists currently visit the Arctic each year, up from one million in the early 1990s. Longer and warmer summers keep Arctic seas freer of ice floes, so cruise ships can visit places that were once inaccessible. The annual number of visitors to Svalbard has surged by 33% in the period 2002 to 2006 to about 80 000 (Naik, 2007). At the beginning of the 21st century, tourism has clearly become a major engine of growth for the Alaskan economy. According to the Alaska Tourism Satellite Account (Anon, 2004), travel and tourism's economic contribution in Alaska reached USD 1.6 billion in 2002. This amount – representing sales net of related imports into the state – contributed 5.6% to Alaska Gross State Product (GSP). This includes the direct and indirect effects of all travel and tourism expenditure, but not induced (multiplier) effects. Travel and tourism's core industry – that is, only the direct impact of end-providers of goods and services to travelers – generated USD 856 million in local value added in 2002: 3.0% of Alaska's GSP. Using the core industry definition, travel and tourism was the third largest private sector employer – fourth overall – in the State, with 26 158 direct full-time-equivalent jobs (9.1% of Alaska's total employment) in 2002. Travel and tourism-generated jobs provided USD 579 million in core labor income (benefits and salaries) to Alaska. Adding public sector employment, travel and tourism's economic contribution to employment reached 39 420 full-time-equivalent jobs. Those jobs provided Alaskan workers with USD 1.15 billion in income. Glacier-related tourism is an important part of the economy in several regions of the Arctic, especially in Alaska. Every year hundreds of thousands of people visit coastal Alaska to see the glaciers. Melting of the glaciers in the warming Arctic may be a threat to this sector of the economy.

7.8. New information expected from International Polar Year projects

- Eleven projects focusing on the behavior of Arctic mountain glaciers and ice caps were initiated during the International Polar Year (IPY). They contributed to several IPY projects including GLACIODYN and 'State and Fate of the Cryosphere'.
- New regional surveys of glacier changes, which contribute to the IPY program 'State and Fate of the Arctic Cryosphere', have been conducted in Alaska, the Yukon, Labrador, Iceland, the Russian Arctic Islands, and northeastern Siberia. Surveys in Alaska and Iceland involve new mapping of glacier surface topography with LIDAR and laser altimetry. Satellite laser altimeter data from ICESat have also been used extensively in these surveys. The Yukon surveys reveal a 22% reduction in glacier area since the 1957/1958 International Geophysical Year.
- GLACIODYN projects focus on understanding and modeling the role of changing ice dynamics in the response to climate change of polythermal and calving tidewater glaciers. Target glaciers for this work are Columbia and McCall Glaciers in Alaska, Belcher Glacier on Devon Island, Canada, and Austfonna and Vestfonna in Nordaustlandet, Svalbard.
- A Russian IPY project has focused on the formation, dynamics, and decay of icebergs in the western sector of the Russian Arctic, a topic of particular interest given the increasing amount of shipping and offshore resource exploration activity in the region.

7.8.1. Alaska

Three U.S. IPY projects will contribute to knowledge of Arctic glacier changes and mechanisms of change. As part of IPY-37 GLACIODYN, there is a continuation of the 30-year record of observations on the retreat of Columbia Glacier, Alaska, which at present is the largest single Alaskan contributor to sea level change (at 3.3 Gt/y from 1982 to 2001, and 6.6 Gt/y around 2001). The ongoing work at Columbia Glacier includes re-analysis and archival of the 145-mission, 30-year sequence of aerial photography of the glacier (the longest and most detailed record of its kind ever acquired), a continuation of this aerial photogrammetric record, and a variety of additional measurements and analyses including photogrammetric feature-tracking, terrestrial time-lapse photogrammetry, airborne radar, GPS surveying, seismic observations of calving, and meteorological monitoring. Collaborative work on Svalbard with Polish colleagues is included in this project. The results will add to knowledge of the physics of calving and tidewater glaciers, and improve the ability to predict future marine-ending glacier behavior not only on Arctic glaciers and ice caps, but on marine-ending glaciers everywhere, including the outlet glaciers of Greenland and Antarctica. Work under this funding is in progress and research supported in part by this funding is presented by O'Neel and Pfeffer (2007).

A second GLACIODYN project aims to investigate the mass balance and ice dynamics of polythermal land-terminating glaciers using McCall Glacier in Arctic Alaska as the primary field site. During the IPY, ice cores to bedrock were extracted at three locations. One complete core and selected parts of the other two cores have been returned for analyses designed to elucidate past climate conditions in the region and establish a history of pollutant deposition. Thermistor strings were installed to bedrock in the drill holes as part of a study of internal accumulation within the glacier and its effects on the glacier's mass balance and thermal regime. A program of stream hydrology measurements was initiated both at the glacier terminus and in the river that drains from the region to the Arctic Ocean (these rivers have never previously been gauged and glaciers currently contribute roughly 90% of their discharge). A full 3D higher-order ice flow model of the glacier is being run using high-resolution bedrock and surface maps. To help understand how representative the McCall Glacier measurements are of a broader region, a new LIDAR digital elevation model covering about 500 glaciers in the eastern Brooks Range was created based on surveys conducted in 2008. It will be used to measure changes in glacier volume in the region.

As part of IPY-105 'State and Fate of the Cryosphere', glacier mass loss across Alaska is assessed using airborne laser altimetry along glacier centerline profiles. Repeat profiles are compared to previous mass change estimates to address the primary question of continued acceleration of glacial wastage.

7.8.2. Arctic Canada

Two Canadian IPY projects will contribute new knowledge about the glaciers in Arctic Canada. One component of the project 'Variability and Change in the Canadian Cryosphere' (a contribution to IPY-105 'State and Fate of the Cryosphere') involves producing a multi-temporal glacier inventory and mapping changes in the extent of glaciers in the Yukon and Labrador since the International Geophysical Year (1958/1959) from aerial photography and satellite imagery. Initial results suggest a reduction of 22% in Yukon glacier area over that period (Barrand and Sharp, 2010) and a 25% reduction in the extent of Labrador glaciers between 2003 and 2007. A second goal of the project is to undertake annual mapping of melt extent, melt duration, melt onset and freeze-up dates, and the end-of-summer distribution of snow and ice facies on all large glaciers and ice caps in the Arctic (including Greenland) using enhanced resolution data from the SeaWinds scatterometer on QuikSCAT. The records extend over ten years, to the end of the 2009 melt season, and will provide new insight into the evolution of summer climate over the ice-covered areas of the Arctic land mass.

A second project is a Canadian contribution to IPY-37 GLACIODYN, which aims to investigate the role of ice dynamics in the response of Arctic glaciers and ice caps to global warming, and improve the ability to predict future changes and their impact on global sea level and fluxes of freshwater to the ocean. The Canadian contribution to GLACIODYN focuses on the Belcher Glacier in the northeastern sector of the Devon Island Ice Cap, Nunavut. This is the fastest flowing outlet from the ice cap and a major source of icebergs. The project involves an intensive field and remote sensing study of the hydrology and dynamics of the glacier linked to the development and validation of a state-of-the-art, high-order coupled model of

ice flow dynamics and glacier hydrology. The model will be used to test hypotheses about the effects of climate warming on meltwater inputs to Arctic outlet glaciers and their impact on ice flow, and to perform simulations of the response of the Belcher Glacier to recent and projected future climate change.

7.8.3. Iceland

A major Icelandic IPY project involves mapping the surface topography of Icelandic ice caps using LIDAR. The results will be used to correct digital photogrammetric maps from the 1990s, allowing an analysis of surface elevation changes of the ice caps over the past one to two decades. The LIDAR mapping is a collaboration between the Icelandic Meteorological Office and the Institute of Earth Sciences of the University of Iceland and is funded by RANNIS (the Icelandic Centre for Research), Landsvirkjun (the National Power Company of Iceland), the Icelandic Public Road Administration, the Reykjavík Energy Environmental and Energy Research Fund, and the National Land Survey of Iceland.

High-resolution LIDAR digital terrain models (DTMs) of most of Hofsjökull, Langjökull, Eiríksjökull, and Snæfellsjökull have been made from surveys conducted during the 2007/2008 IPY. DTMs of Mýrdalsjökull, Eyjafjallajökull, Hofsjökull, and southeastern Vatnajökull were completed in 2010. Mapping will be continued in 2011 to 2013 to produce DTMs of the southern, western, and northern drainage basins of the Vatnajökull ice cap.

The new LIDAR DTMs of proglacial areas and nunataks (isolated peaks protruding through the surface of the ice caps) will be used as a reference for re-analysis of the digital photogrammetric maps of glacier ablation areas in the 1990s, which should reduce systematic bias in these maps. This will make it possible to compare the photogrammetric maps with the LIDAR DTMs and quantify volume changes of the Icelandic ice caps during the past 10 to 20 years.

It is important that the glaciers are accurately mapped now when rapid changes are occurring in response to a warming climate. Accurate DTMs make it possible to pursue various lines of glaciological research, but they also have considerable economic value for the assessment and monitoring of water resources for hydroelectric power generation. Potential uses of the DTMs include modeling of surface mass balance and ice flow, mapping the boundaries of surface and subglacial watersheds, mapping subglacial water flow pathways, and quantifying rates of glacier change and their impact on global sea level. The DTMs are also important for investigations of glacier surges and isostatic uplift due to reductions in ice mass.

7.8.4. Svalbard

During the IPY, two international project consortia worked on glaciology in Svalbard: the IPY Kinnvika project and GLACIODYN. Both studied mass balance and dynamics with a focus on large ice caps. The glaciological part of the Kinnvika project focused on Vestfonna (2500 km^2) and involved studies of accumulation rates, ablation rates, ice temperatures, and ice geometry and measured ice velocity fields. Fixed mass balance stations and automatic weather stations were established across the ice cap. Shallow cores were taken to find the average point mass balance over a few decades. The data provide input to and validation of modeling and remote sensing studies.

The Norwegian part of GLACIODYN focused on Austfonna in Nordaustlandet and on Kongswegen / Kronebreen in northwestern Spitsbergen. GLACIODYN monitored the current status of more than a dozen Arctic ice caps and glaciers with the aim of predicting the future fate of these terrestrial Arctic ice systems. The Austfonna investigations focused on (i) direct surface mass balance measurements; (ii) elevation and volume changes using satellite data, airborne laser profiles, and ground-based GPS measurements; (iii) dynamics of surge, calving and subglacial processes; and (iv) modeling of mass balance and dynamics. Five years of net surface mass balance studies on Austfonna show a slightly negative average balance (-0.1 m w.e./y), but with large interannual variations both in snow accumulation and total ablation. The calving is important (-0.30 m w.e./y) and represents 30% to 40% of the total ablation, similar to values observed on other Arctic ice caps. However, the elevation change measurements on Austfonna show a thickening in the interior of about 0.5 m/y and an increasing thinning closer to the coast of 1 to 2 m/y. A total of 230 km, or about 70%, of the ice cap margin is a calving front, at which all the ice is grounded. The calving front has been retreating at an average rate of 50 m/y since 1990 (Dowdeswell et al., 2008). There seems to be a large dynamic instability, as the measured mass flux is much less than the flux that would be expected given the observed surface mass balance.

On Spitsbergen, a Russian IPY project has used surface radio echo sounding surveys and the analysis of available multi-temporal cartographic, aerial, and satellite-derived data to determine changes in the surface elevation, thickness, ice volume, and internal structure of Fridtjovbreen resulting from a surge event in the 1990s.

7.8.5. Russian Arctic and mountains

The Russian National Program and Implementation Plan for the IPY 2007/2008 includes three projects relevant to glaciers in the Russian Arctic. The project 'Current state of glaciers and ice caps in the Eurasian Arctic' is a Russian contribution to GLACIODYN. The main goal of the project is to study the area changes, mass balance, hydrothermal state, and potential instability of glaciers and ice caps in the Russian Arctic islands and Svalbard. The main fieldwork programs in 2007 and 2008 included airborne and surface radio echo sounding surveys of ice thickness and bedrock and surface topography of ice caps and glaciers, which have been supported by the analysis of spaceborne remote sensing data. On the basis of radio echo sounding surveys of glaciers on Franz Josef Land and Novaya Zemlya and satellite altimetry data, characteristic heights and thicknesses of glacier fronts producing icebergs have been determined, including Glacier No. 1 and the Moscow Ice Cap on Hall Island, the northern part of the glacier complex on George Land (Franz Josef Land), and the glaciers in the Inostrantsev Bay area, Novaya Zemlya. New criteria for the estimation of iceberg hazards from the glaciers of Novaya Zemlya and Franz Josef Land were developed. Franz Josef Land has the greatest potential for regular formation of large icebergs (with thicknesses of up to 150 to 200 m and extents of more than 1 to 2 km) (Kubyshkin et. al., 2009).

A second project, 'Formation, dynamics and decay of icebergs

in the western sector of the Russian Arctic', aimed to collect new data on the formation, distribution, and properties of icebergs in the Barents and Kara Seas and to estimate the current state of outlet glacier fronts in the Russian Arctic archipelagos. During a cruise of R/V *Mikhail Somov* in September 2007, iceberg-producing glaciers on Franz Josef Land, Novaya Zemlya, and some other islands were surveyed. The survey included a wide range of iceberg studies, following their life history from calving to melting and disintegration. The study continued the survey of 2003, when an abnormally large group of icebergs (over 40, the largest of which was over 400 m long and weighed about 3 million tonnes) was discovered in the northeastern part of the Barents Sea, in the vicinity of the *Shtokman* gas-condensate field. The unusual ice conditions allowed R/V *Mikhail Somov* to circumnavigate Franz Josef Land. Helicopter radio echo sounding and aerial photography of glaciers on Prince George Land, Salisbury Island, Luigi and Champ Islands, Hall Island, and Wilczek Land allowed determination of ice thickness in potential iceberg-producing areas. Photogrammetry was used to reconstruct the geometry of glacier fronts and the above-water parts of icebergs. On some glaciers and icebergs, the vertical distribution of ice temperature (down to 20 m depth) and energy balance components of the upper 3 m layer were measured. Observations were made on the surface and in the ice (down to 3 m depth). On some glaciers, markers were installed for measurement of glacier flow. Glaciological studies were supplemented with oceanographic temperature and salinity profiling in the Franz Josef Land straits. Several groups of large tabular icebergs (more than 1 million tonnes) were found not far from their calving areas (Elena Guld Bay on Wilczek Land; the straits between Salisbury, Luigi, and Champ Islands; Geographers' Bay on Prince George Land). The majority of large icebergs were already drifting. Under favorable meteorological conditions, some may drift to the Barents Sea through the deep straits. A similar survey was undertaken of the glaciers of the northern end of the Northern Island of Novaya Zemlya (Buzin et al., 2008). This survey was repeated in autumn 2008.

The project 'Climatic factors in the contemporary evolution of North-Eastern Siberia glaciation' is another Russian contribution to GLACIODYN. This project continued studies of climate–glacier interactions in the poorly explored northeastern Siberia region. This region is of special interest because its climate is influenced both by Atlantic and Pacific air masses. It includes a part of the Northern Hemisphere where climate change (weakening of the Siberian High and warming in recent years) and cryospheric change have been detected. Meteorological parameters were measured along a transect from Magadan to Oymyakon, and in the northern massif of Suntar-Khayata. A study has been completed of glacier change in the region based on modern satellite images and data from the USSR Glacier Inventory. Infrared, visual, and aerial photo surveys were conducted for the Suntar-Khayata glaciers in order to update the Glacier Inventory (Ananicheva and Kapustin, in press b).

7.9. Synthesis of the current state of Arctic mountain glaciers and ice caps

- Mass balance assessments based on field measurements and geodetic surveys indicate a marked increase in the rate of mass loss from Arctic glaciers (especially those in Alaska, the Canadian Arctic, and Greenland) since the mid-1990s. From 1961 to 2004, the mean rate of loss was in the range -91 to -141 Gt/y, but pentadal estimates for the period since 1995 range from -170 to -280 Gt/y.
- The recent increase in mass loss coincides with increases in summer air temperature and reductions in glacier area.
- Recent rates of mass loss are comparable to those from the Greenland Ice Sheet, emphasizing the importance of the Arctic as a whole as a first-order control on the rate of global sea level rise.
- Iceberg calving is an important component of mass loss in Alaska, Arctic Canada, the Russian Arctic, and Svalbard (where it accounts for about 32% of the total mass loss). Individual tidewater glaciers can discharge very large amounts of ice into the ocean (e.g., Columbia Glacier discharged 6.6 Gt/y in 2001). However, temporal changes in this mass loss term are not well known and are not included in projections of future mass and volume change.
- Model-based projections suggest a 12% to 32% reduction in Arctic glacier volume by 2100 due to changes in surface mass balance alone. While some individual glaciers will disappear during this period (and others have already done so), mountain glaciers and ice caps in the Arctic will continue to influence change in global sea level well beyond 2100.
- Meltwater runoff from Arctic glaciers is expected to increase for a few decades at least, but will eventually decrease as the influence on melt production of decreasing glacier area becomes more significant than that of increasing melt rate.

7.9.1. Mountain glacier and ice cap mass balance

Two global mass balance datasets, one based on glaciological mass balance measurements (DM05) (Dyurgerov and Meier, 2005) and the other a combination of both glaciological and geodetic measurements (C09) (Cogley, 2009a), are subsetted to determine a pan-Arctic glacier mass balance. DM05 is a weighted arithmetic mean of single-glacier time series within a set of climatically homogeneous regions, based on data from the World Glacier Monitoring Service (WGMS, which collects standardized observations of change in the mass, volume, area, and length of glaciers worldwide; www.wgms.ch), individual published sources and, in some cases, unpublished measurements. Each time series is weighted by the area of its glacier, and then extrapolated to the climatic region based on estimates of the total glacierized area in each region. Nearly all the glaciological measurements in DM05 are derived from land-terminating glaciers (with larger glaciers underrepresented). C09 uses a nearly identical glaciological dataset to DM05, but also includes geodetic datasets that sample

several tidewater glaciers. For the annual glaciological dataset, C09 uses a spatial interpolation algorithm that fits a polynomial in geographical coordinates (centered successively on each cell of a 1° by 1° grid) to the complete (glaciological plus geodetic) set of single-glacier balances. The measured mass balances are inverse-distance weighted to the center of the grid cell, and the resulting estimate is multiplied by the total glacierized area within that cell. Estimates are averaged over five-year (pentadal) time periods. For the geodetic measurements, which usually span several years, the time-averaged mass balance during the entire period is assumed to represent the annual balances within that period. Interannual variations in the geodetic balances are accounted for by calculating a standard deviation that is a function of the warmest month's air temperature in a given year, interpolated to the glacier equilibrium line. The temperatures used are climatological means from the NCEP/NCAR Reanalysis.

While all the mass balance measurements included in the DM05 and C09 datasets are total mass balances (surface mass balance plus iceberg calving), very few are from calving glaciers. Circumstantial evidence presented by Cogley (2009a) suggests that calving glaciers may currently have more negative mass balances than land-terminating glaciers. If so, underrepresentation of calving glaciers in these datasets may result in underestimation of rates of mass loss in both global mass balance assessments. As yet, it is not understood why rates of mass loss from calving glaciers might be higher than those from land-terminating glaciers and it is not possible to properly quantify the magnitude of the bias in global mass balance assessments that arises from underrepresentation of calving glaciers in the datasets on which these assessments are based. Equally, it is not yet possible to predict how the rate of mass loss by iceberg calving will evolve under different climate change scenarios.

Both estimates derived from measurements (DM05 and C09) show that mass balances were predominantly negative during the period of record, with large interannual variability (Figure 7.29). During the period 1961 to 2004, the average mass balance of Arctic glaciers was -91 Gt/y using the DM05 data and -138.4 Gt/y using the C09 data. Estimates from DM05, based only on glaciological measurements, are more positive than the C09 estimates, possibly because they underrepresent calving glaciers. A third global mass balance estimate, derived from a surface mass balance model driven by climate station data (Hock et al., 2009), suggests that the surface mass balance of Arctic glaciers was -141 ± 65 Gt/y during the period 1961 to 2004 (Figure 7.29). Mass balances were significantly more negative in the period 1995 to 2005 than previously (of the order of -170 to -280 Gt/y according to the C09 dataset). These recent rates of mass loss are comparable in magnitude to recent estimates of mass loss from the Greenland Ice Sheet (see Chapter 8). When the mass loss rates from Arctic mountain glaciers and ice caps are combined with those from the Greenland Ice Sheet, it becomes clear that wastage of glacier ice in the Arctic is a first-order control on the rate of global sea level rise.

The regional contributions that comprise the pan-Arctic mass balance estimate derived using the combined geodetic-glaciological method (C09) are shown in Figure 7.30. This figure also shows the pentadal mean summer (June to August) 700 hPa air temperature for each region from the NCEP/NCAR

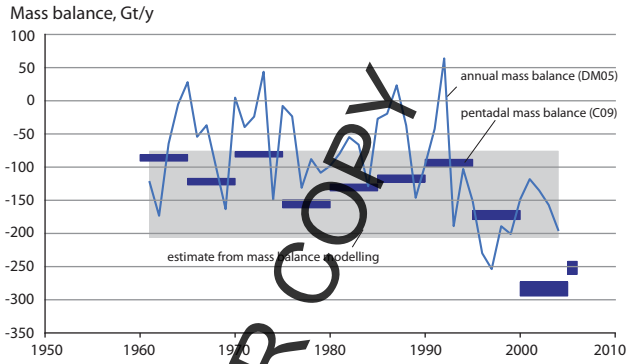


Figure 7.29. Annual mass balance (Dyurgerov and Meier, 2005) and pentadal mass balance (from data compiled by Cogley, 2009a) of Arctic glaciers as outlined in Figure 7.9. Annual estimates are extrapolated from glaciological measurements. Pentadal estimates are derived from a combination of geodetic and glaciological data. The grey box is an estimate from mass balance modeling (Hock et al., 2009) marking the error bounds around the mean estimate. Source: AMAP.

Reanalysis. The time series of estimated contributions from regions with observed records (such as southern Alaska, the Canadian Arctic, Scandinavia, and Iceland) are inversely related to regional summer air temperature variations. However, the time series for some regions with fewer observations (such as the Russian Arctic) are not consistent with local climate fluctuations. In these regions, the reconstructed mass balance records may be overly influenced by regional interpolation from areas with more mass balance data and different temperature histories. In general, the degree of confidence in the regional mass balance reconstructions decreases as the amount of smoothing of observed data required to generate them increases. This suggests that records from such regions need to be treated with caution and that deriving more reliable records for such regions should be a high priority.

Estimates of regional mass balance based on local measurements also illustrate the relative contribution of each region to rising sea level (Figure 7.31) over specific intervals of time. All areas with regional syntheses show mass losses during their periods of measurement, with Alaska and the Canadian Arctic being the dominant contributors to rising sea level. It is important to note that these estimates for different regions are not directly comparable owing to the differences in measurement periods.

7.9.2. Ablation due to iceberg calving

The relative proportions of ablation due to calving and surface mass balance are not well known, but available data suggest that calving plays an important role in the mass balance of glaciers in Alaska, the Canadian Arctic, Svalbard, and the Russian Arctic. There are 54 tidewater glaciers in northwestern North America, some of which are continuing their rapid retreat that commenced about 200 years ago. Among these, Columbia Glacier is undergoing the most rapid retreat exceeding 1 km/y since 2000, discharging an average of 3.3 Gt/y of calved ice during the period 1982 to 2001, with 6.6 Gt/y calved ice in 2001 (O'Neel et al., 2005). In Svalbard, calving losses accounted for about 32% of the overall glacier ablation (Błaszczyk et al., 2009). In Arctic Canada, Svalbard, and Severnaya Zemlya, calving comprises a significant fraction of the total ablation from several large ice caps including Devon Ice Cap and Prince

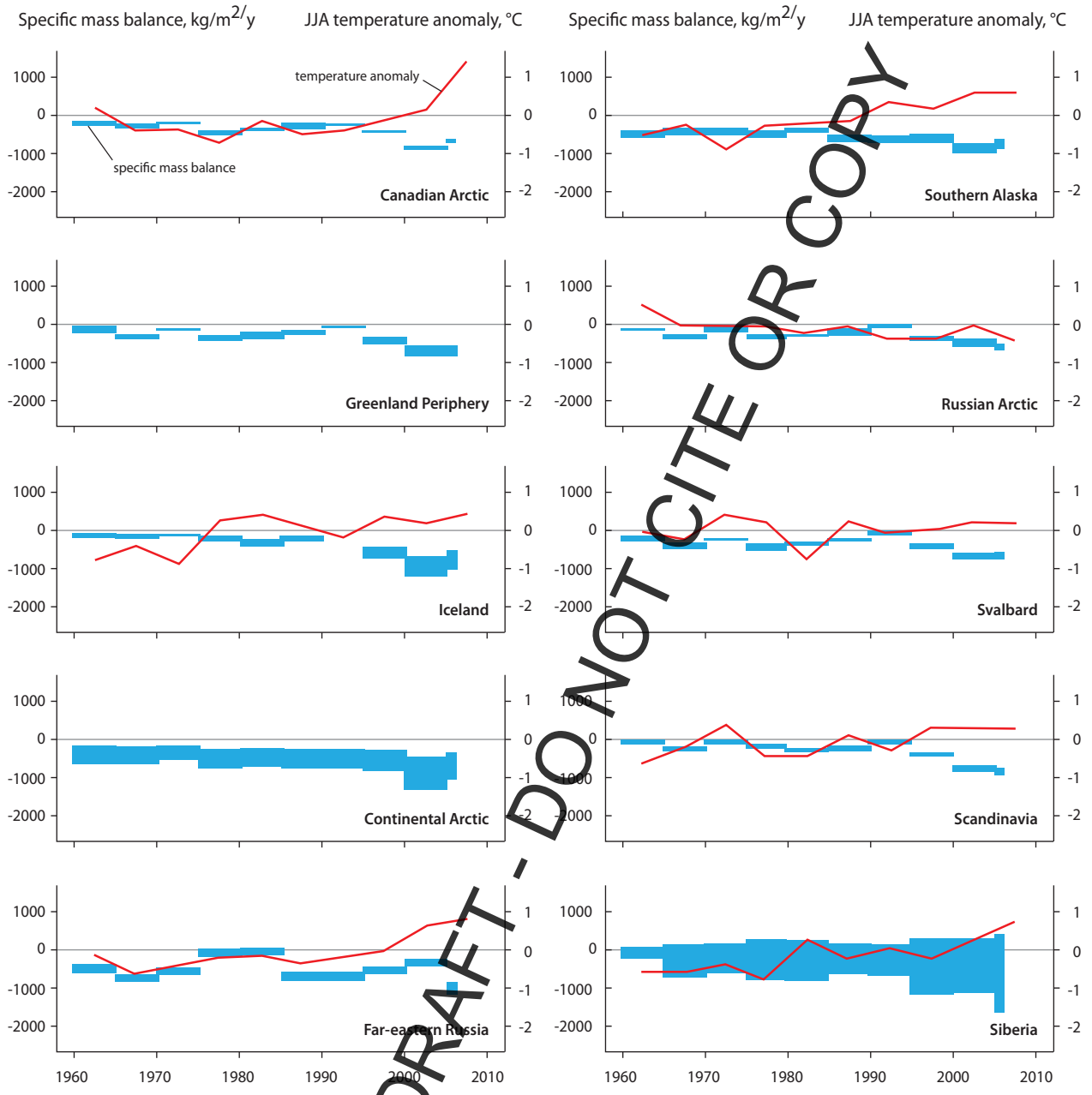


Figure 7.30. Estimates of pentadal mean annual specific mass balance for glacierized regions of the Arctic derived using the combined geodetic-glaciological dataset of Cogley (2009a) and the pentadal mean of annual anomalies in summer (June – August) air temperature at a geopotential height of 700 hPa over each region from the NCEP/NCAR Reanalysis. Temperature data are not plotted for the Greenland Periphery and Continental Arctic regions due to the non-contiguity of these regions. Source: Dan Moore, University of British Columbia.

of Wales Icefield, Austfonna, and Academy of Sciences Ice Cap (Dowdeswell et al., 2002, 2008; Burgess et al., 2005; Williamson et al., 2008). However, ongoing monitoring of calving fluxes is almost non-existent and, as a result, the magnitude of temporal variability and the existence and sign of any trend in calving fluxes are essentially unknown.

7.9.3. Observed trends in ice extent

Observed changes in ice extent over the past half-century are consistent with changes in mass. In nearly all regions, mass losses occurred coincidentally with decreases in glacier area. Although a small number of glaciers have advanced during this time, nearly all of these are tidewater or surge-type glaciers. The area evolution of tidewater and surge-type glaciers depends more on dynamic cycles than on direct climatic forcing, although it may be influenced by regional climate patterns.

7.9.4. Relationship to climate and circulation changes

Arctic air temperatures have increased during the past half-century (Hinzman et al., 2005), which probably explains much of the increase in rates of mass loss observed on Arctic glaciers. Increases in summer air temperature explain the accelerated losses from glaciers in northwestern North America (Arendt et al., 2009) and Arctic Canada (Gardner and Sharp, 2007), but uncertainties remain large due to the difficulties in modeling dynamic losses owing to calving.

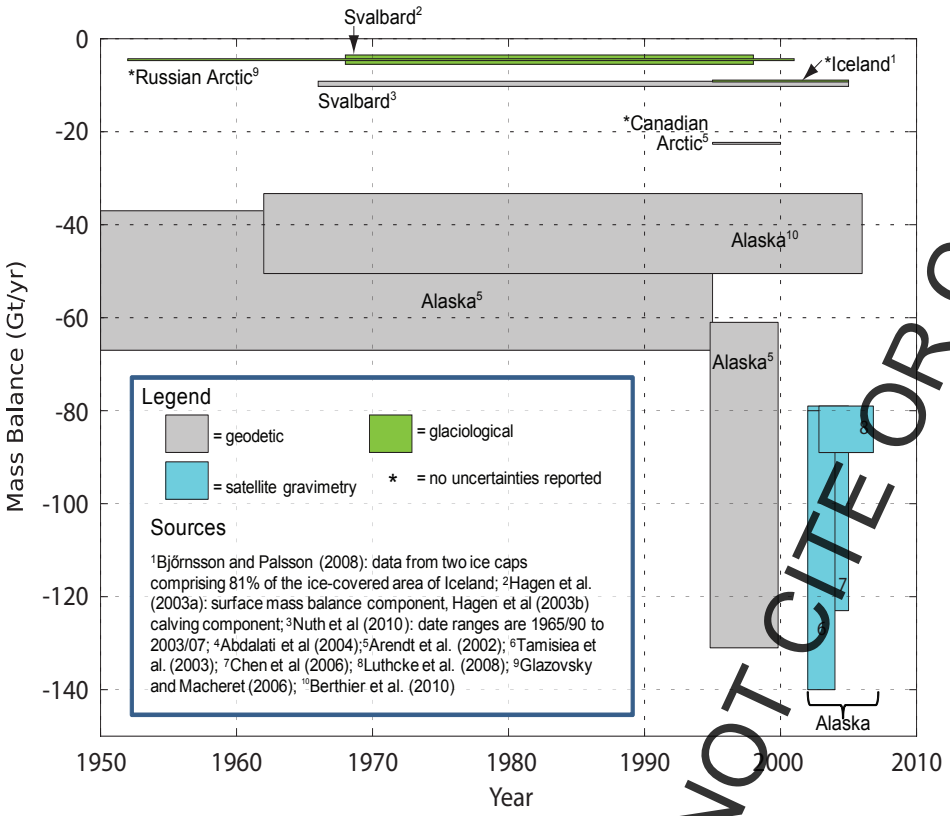


Figure 7.31. Estimates of the mass balance of Arctic glacier regions for which regional assessments have been conducted. The vertical extent of each box describes the range of uncertainty in the mass balance estimates (uncertainty estimates are not available for Iceland or the Canadian or Russian Arctic). Glaciological estimates relate only to the surface mass balance. Source: Anthony Arendt, University of Alaska, and W. Tad Pfeffer, INSTAAR, University of Colorado.

In general, changes in synoptic-scale atmospheric circulation patterns, resulting from variations in sea-surface temperature, polar jet stream position, and other factors, correlate well with Arctic glacier mass balance time series. In northwestern North America, two well-documented 'regime shifts' occurred in 1977 and 1989, during which temperatures abruptly shifted to more positive values and storm tracks changed direction. These shifts are evident in the mass balance time series of glaciers in Alaska (Bitz and Battisti, 1999; Josberger et al., 2007). In the Canadian Arctic, accelerated surface mass loss after 1987 was associated with changes in the position of the July circumpolar vortex. In the Russian Arctic, mass balance changes have been linked to sea-surface temperature variations in the Barents Sea.

7.9.5. Projections of future change

Modeling studies indicate that many individual glaciers may disappear by the end of the 21st century if air temperatures in the Arctic continue to rise as projected in the scenarios used by the suite of climate models employed in the Fourth Assessment of the IPCC (Solomon et al., 2007). The total volume of Arctic glaciers is projected to decline substantially by 2100. Projected reductions are of the order of 12% to 32% of current glacier volume. However, these projections vary greatly from model to model, are sensitive to the choice of climate scenario, and generally do not include mass losses by iceberg calving, which is known to be an important mode of ablation. Nonetheless, it is clear that glaciers in the Arctic are likely to continue to influence changes in global sea level well beyond the end of the 21st century.

7.9.6. Impacts

In many parts of the Arctic, climate warming should cause glacier runoff to increase for a few decades or longer, but glacier retreat will ultimately cause glacier runoff to decline. Runoff decline may already have started in some regions where most glaciers are small.

Changes in glacier runoff will influence freshwater habitat, water supply, hydroelectric power generation, flood and avalanche hazard, estuarine and coastal habitat, and patterns of ocean circulation. Glacier retreat may result in the destabilization and failure of valley-side slopes. Outburst floods from moraine-dammed lakes are likely to decrease in frequency as glaciers retreat further from moraines formed during the Little Ice Age. The frequency and magnitude of outbursts from existing glacier-dammed lakes may also change as ice dams are thinned by ice melting. However, new lakes may form further up-glacier as glacier margins recede down valley-side slopes and melt rates increase.

7.10. Knowledge gaps and recommendations

- The major gaps in knowledge that limit the ability to quantify current rates of glacier wastage in the Arctic and predict future trends are (i) the lack of a complete glacier inventory for the Arctic (including information about glacier location, area, surface topography, and ice thickness); (ii) the limited number of *in situ* mass balance measurements, the strong spatial bias in the distribution of available measurements, and the complete lack of measurements from areas such as the Russian Arctic; and (iii) the lack of ongoing monitoring programs for the iceberg calving component of mass balance, and limited knowledge of the major controls on calving fluxes.
- A coordinated effort to fill these knowledge gaps is needed in order to predict more reliably the current and future

contribution of Arctic glaciers to global sea level change and better assess how glacier change will affect regional water resources.

7.10.1. Key gaps in knowledge

Despite progress since the Arctic Climate Impact Assessment (ACIA, 2005), major gaps remain in knowledge about Arctic glaciers and ice caps. The keys gaps are as follows:

- There is a severe lack of basic information about glaciers in the Arctic. Knowledge gaps include the geographical distribution, total area and size distribution of glaciers in some regions, the distribution of area with elevation (hypsometry), and glacier thicknesses and, therefore, volume (Figure 7.32).
- Glaciological mass balance measurements are limited to a small number of Arctic glaciers (50 in the 2000 to 2004 pentad, of which 30 were in Scandinavia), with larger glaciers and calving glaciers underrepresented. There have been no mass balance measurements conducted in Arctic Russia since 1990, and all but one measurement series from that region ended before 1980. In Arctic Canada, the number of measurement sites peaked at 13 in the early 1970s and is now only five. Given that the uncertainties associated with estimates of regional mass balance increase substantially as the number of local measurements decreases, there is a pressing need to continue (and expand) *in situ* monitoring programs and to develop alternative well-validated methods for assessing regional-scale glacier mass changes, including remote sensing observations (although a problem with these is the lack of temporal and methodological continuity between sensors and satellite missions) and numerical modeling.
- Observations, process understanding, and modeling capabilities related to ablation by calving are severely limited, introducing large uncertainties into current and future mass balance estimates.

7.10.2. Basic observations

Basic information about the geometry and distribution of glaciers is required to assess and model current and future mass changes and contributions of Arctic glaciers to changes in sea level, but this information is currently incomplete for the Arctic. As it is unlikely that direct volume measurements on regional scales will become available in the near future, volume estimates will continue to rely on volume-area scaling techniques. Nonetheless, more direct volume measurements, for example by airborne radio echo sounding, are required to validate scaling techniques.

Completion of high-quality Arctic-wide inventories of glacier areas and size distributions from recent optical satellite imagery is also essential to improve Arctic glacier mass balance assessments. It is important to have regional inventories based on glacier regions; regions should not be bisected by political borders.

There is need for an accurate pan-Arctic inventory of surge-type glaciers and a better understanding of surge recurrence intervals, to provide better estimates of the drawdown of ice by these unusual glaciers. The fraction of mountain glacier and ice

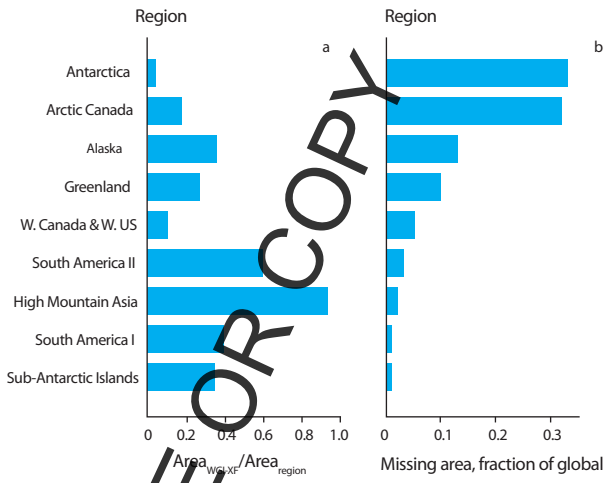


Figure 7.32. (a) Ratio between the area of Extended World Glacier Inventory (WGI-XF) glaciers (Cogley, 2009b), AWGI-XF, and regional glacier area, A_{region}, for the nine glacierized regions that have incomplete glacier inventories. (b) Missing area in WGI-XF as a fraction of missing global glacierized area. While "a" shows the degree of completeness of each regional inventory, "b" indicates how much the different regions contribute to the global missing area. South America I refers to Bolivia, Ecuador and Peru, while South America II refers to Argentina and Chile. Source: after Radić and Hock (2010).

up area drained by marine outlets, and the size and discharge of the marine outlets (including the temporal variability in discharge), must be known if present-day estimates and future forecasts of mass loss by iceberg calving are to be improved.

7.10.3. Mass balance measurements, proxies, and modeling

Most of the uncertainties associated with mass balance estimates arise from the limited number of *in situ* observations. Because the number and distribution of *in situ* observations are unlikely to increase in the short term, there is a need to use remote sensing data to monitor ongoing changes in glacier geometry, surface mass balance, dynamics, and iceberg production. In particular, the potential for estimating mass change of Arctic ice caps using satellite gravimetry (GRACE) and repeat-track altimetry data from the Geoscience Laser Altimeter System on ICESat and its successor needs to be explored more fully. The successful launch in April 2010 of the European Space Agency's CryoSat-2 satellite will create opportunities for monitoring changes in the thickness of the larger ice masses in the Arctic.

Most indices of glacier change at regional scales can only be derived from remote sensing sources, so it is important to continue to develop and maintain long and continuous time series of remote sensing data. The development of regional-scale indicators of mass balance is currently limited by the low spatial resolution of much readily available remote sensing data. This seriously impedes the ability to map the properties of smaller ice bodies containing a significant fraction of the total glacier mass, resolve local features, and properly interpret and validate interpretations using sub-pixel *in situ* data. This is particularly the case with mapping melt duration and identifying the limits of snow and ice facies (especially the superimposed ice zone) from active microwave data that are freely available at the pan-Arctic scale. The accuracy of the results of facies mapping

affects the accuracy of secondary indices (e.g., equilibrium-line altitude and accumulation area ratio) derived from facies maps. The ability to monitor regional-scale mass balance proxies at high spatial and temporal resolution is dependent on the availability of microwave scatterometer data and enhanced resolution products (e.g., QuikSCAT SIR). It is crucial for the construction of long-term records that these data sources be replaced following the recent demise of QuikSCAT.

Uncertainties in nearly all mass balance methods are poorly quantified and rarely reported, although the latter problem is abating steadily. Often it is not clear whether reported mass balances include internal accumulation or calving, both of which can be major components of the mass balance; this makes direct comparison of different measurements problematic. The effects of internal accumulation and superimposed ice formation on glaciological mass balance measurements need to be quantified and these processes need to be included in mass balance modeling. Rigorous identification of sources and magnitudes of error is necessary through intercomparison of methods. Methods for combining mass balance datasets derived from different sampling techniques and representing different time periods and regions need to be expanded and improved to deliver firmer estimates of mass balance and contributions to sea level rise for all Arctic ice masses. The number of glaciologists with expertise in mass balance compilation and integration is also declining and a new generation of glaciologists with these skills must be trained soon.

Given the limited number of *in situ* observations, modeling will continue to play an important and growing role in the assessment of ongoing change in surface mass balance as well as in the projection of future changes. Regional-scale modeling remains a challenge because of the difficulties involved in downscaling climate re-analysis data and climate model results to the complex topography and the heterogeneous climatic conditions of mountain glacier and ice cap environments. These problems are exacerbated by the limited availability of ground measurements made on glaciers to validate the results of downscaling. Projections of future mass balance are also challenging owing to the magnitude of the uncertainty in climate projections for the Arctic that results both from the choice of forcing scenarios used to drive the climate models and large inter-model differences in climate response to the same forcing scenario.

7.10.4. Ice dynamics and ablation by iceberg calving

Although, with the possible exception of internal accumulation (but see Reeh et al., 2005 and Reeh, 2008 for recent progress), processes related to surface mass balance are relatively well understood, there is a severe gap in understanding and modeling of mass loss by calving. Glaciers that terminate in water can have dynamic instabilities that result in rapid mass losses. However, for glacier calving and glacier-marine interactions, there is a lack of both the physical understanding of critical processes and the observations necessary to validate the mathematical representation of these processes. There is also a lack of direct measurements of calving flux and its variability, and of the roles that glacier-marine interactions, bathymetry, ocean temperature, and terminus geometry and dynamics play in determining rates and patterns of glacier retreat and advance. Another area where understanding is limited involves the relationships between summer melt, the development of meltwater drainage systems on, in, and under glaciers, the delivery of water to glacier beds, and the subsequent effects on glacier sliding.

Such information would provide constraints on estimates of total dynamic mass losses and would improve modeling results. There appear to be fundamental differences between calving from tidewater glaciers that terminate in the ocean and calving from those that terminate in freshwater, but these remain poorly understood. Ablation by calving has generally been neglected in predictions of future mass balances. Hence, there is an urgent need to incorporate suitable parameterizations in Arctic-wide mass balance modeling. Since approaches developed for modeling individual glaciers are not yet suitable for application at regional scales due to the paucity of constraining observational data and the largely unresolved nature and complexity of the processes involved, there is a need to explore simpler empirical parameterizations. Given that iceberg keel plowing of sea-floor sediments is a major hazard to underwater engineering structures, there is very little knowledge of the size-frequency distribution of iceberg keel depths or the sources of large icebergs.

7.10.5. Impacts

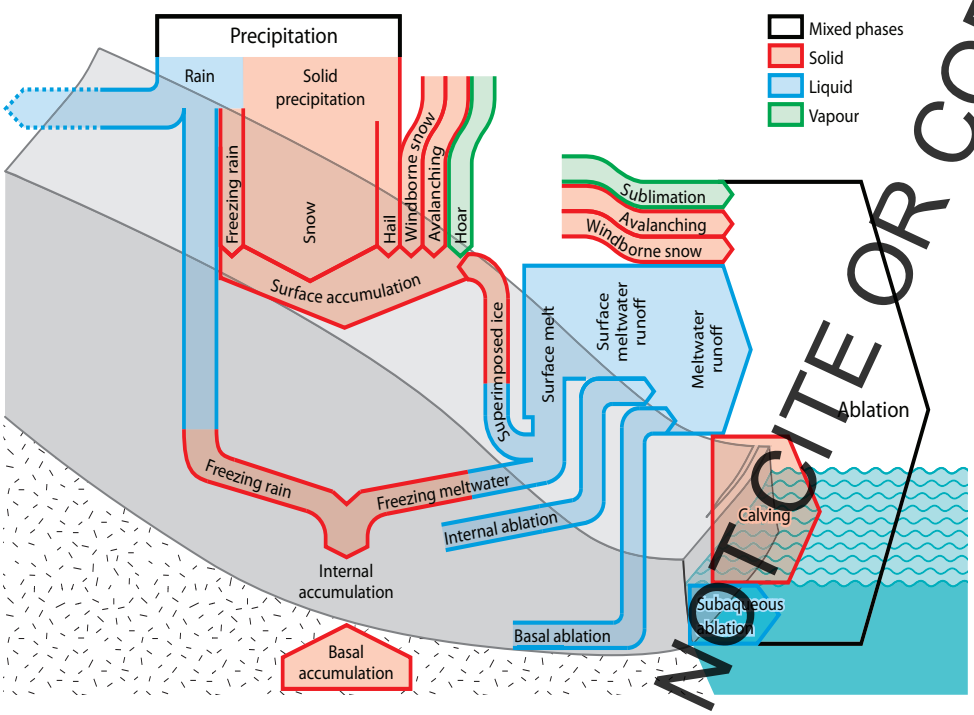
There is a general understanding that climate warming should cause an initial increase in glacier runoff, followed by a decreasing trend. However, the magnitude of peak runoff and the timing of the shift from increasing to decreasing runoff are not known for most watersheds. Indeed, it is possible that some glaciers are currently in the phase of declining runoff, although the total stored ice volume is large enough that an Arctic-wide decline in loss will not occur in the near future. This knowledge gap is particularly critical because changes in glacier runoff control many other impacts, including those on water quality, stream channel dynamics, and circulation in estuaries and coastal waters. There is a need for further simulations of the transient response of glacier geometry and runoff, and their application to a range of drainage basins and climate scenarios, to understand better the sensitivity of runoff changes to climate forcing and its regional variability.

There is a reasonable consensus regarding the direction of stream temperature response to changing glacier runoff and glacier retreat. However, the magnitudes are uncertain. There is less certainty regarding the changes in biogeochemistry and sediment yields. Further process-based research and model development are required to address these gaps. Without predictions of these water quality responses to future climate-glacier change it is not possible to make reasonable projections of the impacts on aquatic ecosystems.

Finally, the impact of climate change on the magnitude and frequency of glacier hazards needs to be understood with a considerably greater degree of confidence to be useful.

Appendix 7.1 Glossary of glaciological terms

Definitions follow those proposed by Cogley et al. (2011).



Caption

Accumulation area ratio

The *accumulation area ratio* (AAR), often expressed as a percentage, is the ratio of the area of the accumulation area to the area of the glacier. The AAR is bounded between 0.0 and 1.0 and its value often correlates well with the surface mass balance. The likelihood that mass balance will be positive increases as the AAR approaches 1.0.

Calving

Calving involves the breaking off of discrete pieces of ice from a glacier terminus into a lake or marine waters, producing icebergs. Calving of outlet glaciers is responsible for roughly half of the mass loss of the Greenland Ice Sheet (see Chapter 8), and can account for a significant part of the mass lost from glaciers in the Arctic.

Calving flux

Calving flux is the *mass flux by calving from a glacier terminus*, with dimension mass per unit time [M/T]. The calving flux can be determined from ice thickness and velocity measurements close to the terminus and the change in ice mass due to terminus advance or retreat.

Equilibrium line

The equilibrium line is defined by the set of points on the surface of the glacier where the surface mass balance is zero at the end of the melt season. The equilibrium line separates the *accumulation area* (where the annual mass balance is positive) from the *ablation area* (where the annual mass balance is negative). It coincides with the snowline (defined as the set of points on a glacier forming the lower boundary of the snow-covered area) only if all mass exchange occurs at the surface of the glacier and there is no superimposed ice. Where superimposed ice forms, the equilibrium line marks the lower limit of superimposed ice formation and occurs at a lower elevation than the snowline.

Equilibrium-line altitude

The spatially averaged altitude of the *equilibrium line* is known as the equilibrium-line altitude (ELA). The ELA is generally determined by fitting a curve to data representing *surface mass balance* as a function of altitude, although it may also be defined by direct visual observation. The ELA often correlates well with the surface mass balance.

Glacier surge

A glacier surge represents anomalously fast glacier motion (10 to 100 times non-surging speeds) sustained for a period of months to years. Surges commonly alternate with slower motion lasting for periods of years to centuries during which the glacier thickens in its upper reaches and thins in its lower reaches. Surging transfers most of the accumulated reservoir of ice from high on the glacier to its lowest reaches, and often extends the glacier terminus by significant distances (kilometres). Surges recur at fairly consistent, glacier-specific intervals. Surge-type glaciers may end on land or in the ocean and are clustered geographically for reasons that are not well understood. Surges seem to be related to changes in the subglacial hydrological regime and not primarily to climatic fluctuations. They are controlled in some cases by thermal processes ('Svalbard-type surging', characterized by long quiescent periods) and

in other cases by hydraulic processes ('Alaskan-type surging' characterized by shorter quiescent periods and higher velocities relative to Svalbard-type surges). While the processes involved in surging are understood in broad terms, many details of their behavior remain elusive (Raymond, 1987).

Internal accumulation

Internal accumulation is the refreezing of surface meltwater (or freezing of rain) that is in transit through the glacier that otherwise would have left the glacier as runoff. Water that freezes within the current year's snowpack is not counted as internal accumulation. Internal accumulation may be regarded as simply redistributing mass within the glacier. On many Arctic glaciers, internal accumulation is a major component in the mass balance and failure to account for it results in a measurement bias towards overestimation of mass loss. Typically, internal accumulation is not detected by traditional surface mass balance measurements.

Mass balance

The mass balance is the change in the mass of a glacier or ice body over a stated span of time, often expressed as a rate. Mass gain (accumulation) mainly occurs by snowfall. On some glaciers deposition of hoar, gain of blowing and drifting snow, avalanching, and basal freeze-on (usually beneath floating ice) are significant processes of accumulation. The glacier loses mass (ablation) mainly by melting and iceberg calving. On some glaciers, sublimation, evaporation, loss of blowing and drifting snow, and ice avalanching are significant processes of ablation. In some cases, a fraction of the meltwater produced (plus rainfall) can refreeze, either within snow or firn (see *internal accumulation*) or on the ice surface (see *superimposed ice*), and does not contribute to mass loss from the glacier. These processes are summarized in Figure 7.8. There are several ways to measure mass balance; these are described in Section 7.3.2.1.

The principal mass balance components are surface accumulation (C_{sfc}, usually mostly snowfall and superimposed ice formation), surface ablation (A_{sfc}, usually mostly melt), internal accumulation C_i, internal ablation A_i, basal accumulation C_b, and basal ablation A_b, and (for ocean- or lake-terminating glaciers) iceberg calving D, so the glacier-wide mass balance $\dot{M}$ expressed as a rate is given by:

$$\dot{M} = \dot{C}_{sfc} + \dot{A}_{sfc} + \dot{C}_i + \dot{A}_i + \dot{C}_b + \dot{A}_b + \dot{D}$$

All terms have the dimension mass per unit time, [M/T]. Surface accumulation and ablation tend to be major components on most land-terminating Arctic glaciers. Internal accumulation is also important on the Greenland Ice Sheet and many polythermal/cold glaciers. Basal ablation can be a dominant component of the mass balance of floating tongues of marine-terminating glaciers, while basal accumulation and internal ablation are generally negligible.

The sum $C_{sfc} + A_{sfc}$ defines the **surface mass balance,** B_{sfc} or SMB, the latter symbol often being used in ice sheet studies. In estimates of ice sheet mass balance, the term has often been extended to include internal accumulation. In addition to snowfall and superimposed ice, surface accumulation also includes deposition of hoar, freezing rain, solid precipitation in forms other than snow (e.g., hail), and snow transported by wind or avalanches.

Mass balance units

The principal dimension for mass balance is mass per unit time [M/T]. Ice sheet balances are often expressed in Gt/y (1 Gt = 10^{12} kg). However, the mass balance, especially of smaller ice bodies, is usually treated as a rate of change of mass per unit area, in which case it is referred to as *specific mass balance* and its dimension becomes [M/L²/T], for example, kg/m²/y, which is numerically identical to the millimetre water equivalent, mm w.e./y. While mass units (e.g., kg or Gt) are useful for hydrological or sea level calculations, specific units (e.g., kg/m², mm w.e.) are needed for comparisons of glaciers of different sizes, for example, when investigating glacier-climate interactions. Mass balances, in particular for larger ice bodies, are often expressed in **sea level equivalent** (SLE), which is the change in mean global sea level that would result if glacier mass were added or removed from the ocean. SLE in mm/y is often simply obtained by

$$SLE = \frac{0.001\dot{M}\,A_{glacier}}{\rho_w A_{ocean}}$$

where the glacier mass change $\dot{M}$ is in kg/m²/y and the areas of the glacier ($A_{glacier}$) and the ocean (A_{ocean}) are in m²; ρ_w is the density of freshwater (1000 kg/m³).

Superimposed ice

Ice accumulated on top of glacier ice by the refreezing of rain or meltwater produced during the current mass balance year is known as superimposed ice. It is different from internal accumulation, which represents refreezing within firn below the summer surface formed at the end of the summer in the previous mass balance year. Measurement of superimposed ice accumulation requires special attention in conventional mass balance programs.

Tidewater glacier

A tidewater glacier terminates in the sea, with its terminus either floating or grounded below sea level. Tidewater glaciers can undergo a periodic growth/shrinkage instability with long periods (centuries) of slower motion and slow advance alternating with shorter periods (decades) of rapid motion, high calving flux, and rapid (~1 km/y) retreat (Meier and Post, 1987).

Termed *tidewater instability*, this cyclic behavior has been documented most extensively in Alaska (Post, 1975), although it has been observed in many other regions, and the present retreat of certain Greenland Ice Sheet outlet glaciers appears to operate in a nearly identical fashion (Howat et al., 2005). During rapid retreat, a glacier can thin far faster (by loss of the marine-grounded tongue plus drawdown of an upstream basin by rapid flow) than can be accomplished by surface mass balance alone. In its advancing phase, a tidewater terminus is stabilized by back-stress against a push moraine at its advancing margin. A tidewater glacier in an advanced position can switch

abruptly (~years) to an unstable phase of thinning in its ocean-grounded reach, fast flow, and rapid calving which results in terminus retreat. Once initiated, fast flow and retreat can continue until the glacier's terminus retreats to a point where the glacier bed rises above sea level.

The causes of unstable retreat are not completely understood, but involve high basal water pressures tied to the depth of the glacier bed below sea level, rapid sliding, vertical thinning, and positive feedback between thinning and fast flow. The processes governing the size and rate of production of icebergs are extremely poorly known. Climate-induced thinning can play a critical role in initiating unstable retreat, but once initiated, unstable retreat is modulated by channel geometry and englacial and basal hydrology and is essentially independent of climate and surface mass balance.

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8. The Greenland Ice Sheet in a Changing Climate

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Key Findings

- At present, Greenland is warming at more than twice the global average due to the polar amplification at high Arctic latitudes. Annual Arctic temperatures are projected to be 2.8 to 7.8 °C higher in 2100.

- Mass balance estimates reveal that the Greenland Ice Sheet is losing ice at an increasing rate. The mass loss has increased from 50 (±50) Gt/y in 1995–2000 to 205 (±50) Gt/y in 2005–2006.

- A slight increment of accumulated mass is more than counterbalanced by the increased loss of mass by melting and ice discharge to the ocean. Runoff and ice discharge have both increased, each with more than 100 Gt of mass lost each year since 1990.

- The discharge occurs mainly through an increase in the velocity of the fast-flowing glaciers that terminate on ocean water in the fjords. Many fast-flowing glaciers have increased their velocity – some as much as doubling their velocity – and the glaciers reacting are observed further and further northward along the west coast of Greenland with global warming.

- The increased velocity from the fast-flowing glaciers appears to be related to warming of the ocean water.

- The present loss of mass (205 Gt/y) from the Greenland Ice Sheet if unchanged will cause a 5 cm global mean sea level rise by 2100. Models predict further loss of mass and increased discharge with the ongoing warming and 10 to 19 cm of eustatic sea level rise from Greenland Ice Sheet can be expected by 2100. However, the models still suffer from a very incomplete understanding of atmosphere-ocean-ice interactions. The maximum projected loss of mass from the Greenland Ice sheet by 2100 is believed to result in a 40 cm eustatic sea level rise.

- Since 2003, the global contribution to sea level rise from the ice sheets and glaciers has increased from 50% to 75–85%. In the Arctic, mass loss from the Greenland Ice Sheet is very likely to exceed the contributions from the Arctic mountain glaciers and ice caps and from thermal expansion of the sea water during the next century.

- The increased meltwater runoff and iceberg discharge are reducing the salinity and density of the recipient marine surface water. These changes result in stronger gradients between the fresh surface water and the underlying saline water, as a result boosting circulation and increasing the transport of nutrients, which has a strong impact on the primary production of the affected marine areas.

- Planned hydroelectric power plants rely solely on runoff from the Greenland Ice Sheet and retreat of the ice sheet needs to be understood to be able to construct reliable and economic hydropower plants.

- Retreat of the Greenland Ice Sheet will open areas for mining, but will result in longer walking distances for hunters.

- Tourism in relation to the accelerating and retreating ice streams, especially at the UNESCO World Heritage site at Ilulissat Ice Fjord, is increasing and plays an important role in first hand witness of climate change.

Summary

The Greenland Ice Sheet is a massive ice cap nearly 3000 m thick. It is the largest body of freshwater ice in the Northern Hemisphere. There are 2.93 million km^3 of ice locked up in the Greenland Ice Sheet. Ice is lost from ice caps either at the surface, where it is melted by warm air and winds, or from the edge, where it breaks off as chunks of solid ice or flows into the ocean as meltwater. Measured rates of both types of loss from the Greenland Ice Sheet have increased.

Most ice comes off the Greenland Ice Sheet in a series of fast-flowing glaciers that discharge to the ocean through fjords along the west coast of Greenland. Studies show that these have increased their rate of flow – some doubled their speed between 1995 and 2000. The trigger for these changes is thought to have been warming of the ocean water that is in contact with the outflowing end of the glaciers. One well-known glacier, the one terminating in the Ilulissat Ice Fjord (Sermeq Kujalleq in Greenlandic), which drains ice from around 7% of the entire ice sheet, has doubled its flow rate and the ice front has retreated by 15 km in the past eight years. These changes are dramatic and the glacier has become a popular tourist destination. Estimates of the amount of ice leaving the Greenland Ice Sheet as solid icebergs have risen from up to 320 gigatonnes per year (Gt/yr) in 1995 to 421 Gt/y in 2005, a jump of 30% in 10 years.

Results published in 2007 indicated that the area of the Greenland Ice Sheet's surface that experiences some melting each year had been steadily increasing since satellite observations began. The area of surface melting is not a direct measure of the volume of water being lost from the ice sheet (called 'runoff'), but both the area and the number of days of surface melting are strongly related to the amount of ice lost.

To calculate the net loss of ice from the Greenland Ice Sheet, these measures of ice loss are offset against ice gains caused by snow falling onto the ice sheet's surface. The height of the Greenland Ice Sheet at its centre, measured from satellites, has risen slightly recently – probably reflecting an increase in snowfall – but this increase is more than outweighed by the increase in loss of ice due to melting and ice discharge.

When mass budgets were first calculated for the Greenland Ice Sheet in the 1960s, the ice sheet appeared to be in balance. Those early measurements were very uncertain, but they showed similar amounts of ice being added and lost each year. The latest measurements tell a very different story, with large net losses of ice every year. The net loss of ice has increased from an average estimate of 50 (± 50) Gt/year from 1995 to 2000 to 200 (± 50) Gt/year from 2004-2008. The current rate of loss (~200 Gt/yr) represents enough water to supply more than one billion city-dwellers.

8.1. Introduction

- The Greenland Ice Sheet is the second largest body of ice on Earth and its mass is equivalent to a global mean sea level rise of about 7 m.
- During the last million years, sea level has ranged from being over 100 m lower than at present during glacial periods to up to 20 m higher during interglacial periods.
- The warming during the past decade in the Arctic is more than twice the global average.
- Annual temperatures over the Arctic (i.e., all areas north of 60° N) are projected to be 2.8 to 7.8 °C higher in 2100 and the mass loss from the Greenland Ice Sheet is projected to increase significantly.

8.1.1. Evolution of the Greenland Ice Sheet

The Greenland Ice Sheet (Figure 8.1) is the largest body of ice in the Northern Hemisphere and, globally, is dwarfed only by the Antarctic Ice Sheet. The entire mass of the Greenland Ice Sheet has a volume of 2.93 million km^3 of ice, equivalent to a global mean sea level rise of 7 m (Table 8.1).

Table 8.1. Total volume of the ice bodies on Earth and the equivalent global sea level change if the ice bodies were to melt completely.

	Volume, million km^3	Equivalent sea level rise, m	Source
Antarctic Ice Sheet	25.4	57	Lythe et al., 2001
Greenland Ice Sheet	2.93	7.5	Bamber et al., 2001
All remaining glaciers and ice caps	0.051 – 0.133	0.15 – 0.37	Lemke et al., 2007

This section presents recent scientific results, focusing specifically on the impacts of contemporary changes in the Greenland Ice Sheet. Although many reports on climate change have addressed impacts resulting from the evolution of the ice sheet, none provide a thorough and up-to-date assessment of the importance of the Greenland Ice Sheet in a changing climate. The importance of the ice sheet as a component of the Arctic cryosphere and the enormous amount of new observations and data provide the background for a better understanding of the evolution of the Greenland Ice Sheet and its potential impacts.

The high latitudes of the Northern Hemisphere are the regions that have experienced the strongest changes in climate since the 1951–1980 reference period (Figure 8.2). While the global annual temperature anomaly for the period 2005–2007 in relation to the 1951–1980 reference period is 0.7 °C, the anomaly over Greenland is 1.5 °C, which is twice the global value. Average surface temperature changes observed in the Arctic are greater than those observed on average over the Antarctic Ice Sheet.

The trend of strongly enhanced changes in surface temperature in the high northern latitudes is not only an issue in terms of the changes observed to date but is also a cause for concern in relation to the predicted increase in future temperatures. The IPCC A1B emissions scenario is projected to result in annual temperatures over the Arctic (i.e., all areas north of 60° N) 2.8 to 7.8 °C higher in 2100 than they were during the 1951–1980 reference period, with winters warming at a greater rate than summers (Meehl et al., 2007). A temperature increase of this magnitude is serious and would be the largest that the Greenland Ice Sheet would have experienced since the previous interglacial period, the Eemian, 130 000 to 115 000 BP (North Greenland Ice Core Project members, 2004).

8.1.2. History of the Greenland Ice Sheet

The global climate has changed enormously over time; there have been periods with glaciations and periods with no ice on Earth. While the Northern Hemisphere experienced ephemeral glaciations from 38 to 4 Myr BP, the onset of extensive glaciations did not occur until 3.3 Myr BP (Lunt et al., 2008).

Over the past three million years, the Greenland Ice Sheet is believed to have waxed and waned in accordance with glacial and interglacial periods consistent with orbital timescale variations in benthic $^{18}O/^{16}O$ ratios (Weidick, 1993; Lisiecki and Raymo, 2005; Weidick and Bennike, 2007) (Figure 8.3). Glacial deposits (i.e., moraines) 20 to 50 km from the present shoreline of Greenland can be used to reconstruct the Greenland Ice Sheet in earlier climatic periods. While the maximum extent of the ice sheet during glacial periods can be observed through studies of moraines and marine sediment cores, the degree to which the Greenland Ice Sheet retreated in earlier warmer periods is not clear.

Observations by Imbrie et al. (1984) indicate that sea level has been higher than at present several times during the past 400 000 years. 130 000 to 115 000 years ago during the Eemian climatic period (MIS5) sea level was 5 to 8 m above present (Kopp et al., 2009; Rohling et al, 2009) and 400 000 years ago during the warm interglacial (MIS20) the highest stand is reconstructed at 20 m above present (Alley et al., 2010).

The 5 to 8 m of additional sea level during the previous interglacial period, the Eemian (MIS5), is of particular relevance for the future evolution of the ice sheet. (Stirling et al., 1998; Lambeck et al., 2002; Kopp et al., 2009).The average air temperature is believed to have been 5 °C warmer than during the 1951–1980 reference period and to have remained so for several thousand years (North Greenland Ice Core Project members, 2004). The reaction of the Greenland Ice Sheet during the Eemian can to some degree be seen as an analogue to the evolution of the ice sheet in future warming scenarios.

Model reconstructions of the ice sheet during the Eemian show very reduced ice volumes (Figure 8.3), especially in southern Greenland (Letréguilly et al., 1991; Cuffey and Marshall, 2000; Otto-Bliesner et al.,

2006), predicting a sea level contribution of 1 to 5 m from the Greenland Ice Sheet. Findings of Eemian ice in several Greenland ice cores and the presence of older ice at the base of the southern Greenland ice core drill site, DYE 3 (North Greenland Ice Core Project members, 2004; Willerslev et al., 2007; de Vernal and Hillaire-Marcel, 2008) support volume changes equivalent to a 1 to 3 m rise in sea level.

During the last glacial maximum, 25 000 to 14 000 BP, annual temperatures over Greenland are believed to have been 25 °C colder than at present (Cuffey et al., 1994; Cuffey and Clow, 1997; Dahl-Jensen et al., 1998), sea level to have been 120 m lower than at present (Lambeck et al., 2002), and the volume of the Greenland Ice Sheet to have been 140% higher than at present (Huybrechts, 2002; Lambeck et al., 2002), corresponding to a global average sea level lowering contribution of 3 m (Fleming and Lambeck, 2004).

Throughout the last glacial period, the climate was very unstable with 25 rapid climate changes, termed 'interstadials' or 'Dansgaard-Oeschger events'. The North Atlantic region and the Greenland Ice Sheet experienced rapid temperature increases of 10 to 15 °C occurring over a few decades (North Greenland Ice Core Project members, 2004; Landais et al., 2005; Steffensen et al., 2008) together with sea level increases of 5 to 20 m (Siddall et al., 2003). The rapid changes are believed to have been connected to enormous surges of ice into the ocean, especially from the Laurentide Ice Sheet over North America and changing strength of the ocean circulation in the Atlantic Ocean (Flückiger et al., 2006). After the abrupt warmings during the Dansgaard-Oeschger events, surface temperatures gradually cooled over the subsequent 1000 to 5000 years, before the next rapid warming occurred. The contribution of the Greenland Ice Sheet to sea level during these rapid warming events is not known. Ice core data show that glacial accumulation over the central parts of the Greenland Ice Sheet was less than 50% of the present accumulation, showing a strong correlation between surface temperature and precipitation (Cuffey and Clow, 1997; Johnsen et al., 2001). The 25 rapid climate changes were also followed by strong changes in precipitation rates, apparent from changing annual layer thicknesses in the ice cores (Alley et al., 1993; Andersen et al., 2005; Svensson et al., 2006).

After the glacial period, the climate warmed into the present interglacial period, the Holocene, and the coastal regions around Greenland experienced an isostatic uplift of up to 100 m in response to the retreating ice (Letréguilly et al., 1991; Weidick, 1993; Huybrechts, 2002).

The onset of the Holocene brought the climate directly into a warm period named the 'Climatic Optimum' 5000 to 8000 years ago with Greenland temperatures 2 °C warmer than in the 1951–1980 reference period (Dahl-Jensen et al., 1998). The onset of the present interglacial period with abrupt warming changed the shape and volume of the Greenland Ice Sheet. The modeled volume changes of the ice sheet entering the warm Holocene climatic period differ depending on the modeled extent of the ice sheet during the last glacial maximum, the mass balance applied, and the evolution of the temperature in the ice (Letréguilly et al., 1991; Huybrechts, 1996; Marshall and Cuffey, 2000; Clarke and Marshall, 2002; Vinther et al., 2009). Investigations in the region around the Jakobshavn Isbræ glacier on the west coast of Greenland show that 5000 to 8000 years ago the margin had retreated 10 km further east than at present (Weidick et al., 1990; Weidick and Bennike, 2007).

Following the Climatic Optimum, surface temperature gradually cooled until The Little Ice Age (occurring between 1500 and 1800) with temperatures 1 °C colder than during the 1951–1980 reference period (Dahl-Jensen et al., 1998). Temperatures then increased and the years around 1940 were about 1 °C warmer than in the reference period. In general, temperature change during the present Holocene climate (the past 12 000 years) has been moderate and of the order of 1 to 2 °C, compared to changes of the order of 10 to 20 °C during the last glacial period. The warming of 1.5 °C over the past decades is of the same order as other warming events recorded during the Holocene interglacial period. Due to the relatively small temperature changes recorded during the Holocene the variability in the proxy paleo-data leading to temperature and precipitation records is of the same order as the changes themselves (Bales et al., 2001, 2009; Mosley-Thompson et al., 2001; Andersen et al., 2005; Vinther et al., 2010). The relation between temperature change and change in mass balance during the Holocene is not well established and is further complicated by regional patterns. Decadal-scale trends over the past century are also complicated by the large degree of natural variability (Chylek et al., 2006).

8.1.3. Evolution of the Greenland Ice Sheet in the future

Experience from modeling the past evolution of the Greenland Ice Sheet indicates the need for improvements in order to accurately model the evolution of the ice sheet in a changing climate. Not enough is known about many fundamental parameters, such as ice sheet mass balance, basal processes, and ice-ocean interactions. Observations are made difficult by the vast extent of the ice sheet and the inaccessibility of the ice sheet base and submerged front.

This inadequate knowledge of processes is affecting the ability to predict the evolution of ice sheets and led the Intergovernmental Panel on Climate Change (IPCC) to conclude that:

> *Taken together, the ice sheets of Greenland and Antarctica are very likely shrinking, with Greenland contributing about 0.2 ± 0.1 mm yr^{-1} and Antarctica contributing 0.2 ± 0.35 mm yr^{-1} to sea level rise over the period 1993 to 2003. There is evidence of accelerated loss through 2005. Thickening of high-altitude, cold regions of Greenland and East Antarctica, perhaps from coastal regions of Greenland and West Antarctica in response to increased ice outflow and increased Greenland surface melting*. (Lemke et al., 2007:376).

Since the latest IPCC assessment was published in 2007 (IPCC, 2007), several other compilations have raised concern about the evolution of the big ice masses in response to global warming (Rignot et al., 2008; Steffen et al., 2008; Velicogna, 2009).

There is consensus from observations that the mass loss from the Greenland Ice Sheet has accelerated since 2000. While the contribution from the ice sheet currently is 10–20% of the observed global sea level rise of 3 mm/y, the Greenland Ice Sheet is believed to be capable of reacting more strongly to the warming over the next 100 years with a mass loss that could increase sea level by several tens of centimetres (Rignot et al., 2008; Steffen et al., 2008).

8.2. The Greenland Ice Sheet today

8.2.1. Overview of the Greenland Ice Sheet

- The Greenland Ice Sheet presents a major orographic barrier to Arctic weather systems.
- Long-term temperature records from coastal stations show a warming over Greenland since 1950. Since 1985 in West Greenland, the warming has primarily been driven by higher winter temperatures.
- The basal hydrological system plays an important role in the flow of the Greenland Ice Sheet and is also affected by increasing amounts of surface water drained to the base.

8.2.1.1. Arctic climate and large-scale synoptic observations

Greenland is the largest island in the world with a surface area of 2.2 million km^2, stretching 2600 km from 59.8 to 83.6° N, where – for several months of the year – it is either polar night or continuous daylight. Over 80% of Greenland is covered by a dome of inland ice (Figure 8.4) rising along an average gradient of about 1% from sea level in various regions to over 3300 m along the central spine.

Greenland plays a pivotal role in determining the climate of the Northern Hemisphere because of its size, location, elevation gradient, and mass of water stored in the ice sheet. During summer, Greenland is ideal for monitoring changes in meridional energy transport into the Arctic because it is located in the predominant direction of cyclone flow. Greenland's extreme North Atlantic location and the ice-filled ocean that surrounds it are the principal factors influencing the climate. The northern branch of the Gulf Stream, known as the North Atlantic Current, runs northward along the Norwegian coast into the Arctic Ocean, mixes with cold polar water and returns southward along the east coast of Greenland. Nearly all the water that leaves the Arctic Ocean passes through Fram Strait (Figure 8.4) via the East Greenland Current, eventually wrapping around Cape Farewell and continuing northward along the west coast of Greenland for several hundred kilometres. A similar, counterclockwise gyre is active along the coast of West Greenland with warm water

moving northward until it meets colder polar water flowing through Kennedy Channel. The resultant cold current flows southward along northeastern Canada.

Wind in the lower troposphere is forced to flow along the coasts of Greenland because of the height and size of the ice sheet. As a result, Greenland is an important participant in the exchange of air mass between the southern and northern latitudes of the Northern Hemisphere. Northerly and southerly air flow are about evenly distributed during summer. However, the predominant flow during winter is northerly due to high pressure in the colder regions to the west or northwest. Airflow in the free atmosphere at the 500 hPa level plays a key role in the Greenland climate because it governs the North Atlantic storm track. Generally, airflow at 500 hPa is from the southwest during winter and from the west during summer.

A typical North Atlantic cyclone develops as a wave in the polar front and propagates along the front. Therefore, winter cyclones generally travel along the east coast of the United States tracking along the edge of the Gulf Stream heading northeast. The cyclones often pass south of Greenland continuing toward Iceland and into the Norwegian Sea. However, they can also track more to the north through Davis Strait and into Baffin Bay. Occasionally, a cyclone will split near the southern tip of Greenland producing separate lows that track along both coasts. During summer, lows tend to be less intense but are frequently more northerly and, therefore, often influence summer conditions in western Greenland.

Passing cyclones are generally accompanied by strong winds. However, during periods with no cyclone activity, the wind regimes are determined by local conditions, which usually relate to katabatic flow on the Greenland Ice Sheet, and sea breezes during summer or land breezes during winter near the coasts. Whether the high velocity katabatic winds on the ice sheet propagate down the fjords to the coastal areas largely depends on the temperature of the air mass as it reaches the head of the fjords. If the katabatic air mass is warmer than the air in the fjord due to adiabatic warming it will not be able to replace the air near the top of the fjord where it will be experienced as a warm 'foehn wind'. As a result, it will ride above the colder and denser surface air mass. If the katabatic flow is colder than the air in the fjord, it will replace the local air mass and probably travel all the way to the coast where it will be experienced as an icy downhill wind.

Climatic conditions in the fjords in the absence of cyclonic or katabatic influences are usually characterized by sea breezes during summer and week land breezes during winter due to ocean-land temperature differences. The winds associated with passing cyclones are, generally, very strong and heavily influenced by local topography and direction of the wind relative to the coast. Winds blowing directly toward the coast will often lead to precipitation due to orographic lifting. This scenario can often lead to very strong barrier winds, which blow clockwise relative to the land mass. Strong winds can cause re-deposition of snow accumulation and even wind-generated mass loss to the sea. Another special feature of the Greenland wind regime is that of very rapid change from calm to gale force conditions. These situations often develop as a result of cold Canadian air reaching the eastern coast of Greenland via the ice sheet behind a cyclone tracking to the northeast. The ice sheet topography determines the direction of the cold outflow from the ice sheet, focusing the wind into the low-lying coastal regions. The Greenlandic term for such an event is *pitoraq* (*piteraq* in East Greenland) which is closely related to the word for 'being attacked'. Piteraqs are most common during autumn and winter. Wind speeds typically reach 50 to 80 m/s. On 6 February 1970, the community Tasiilaq, also known as Ammassalik and the most populous community in East Greenland, was hit by a very strong piteraq causing severe damage. Since the beginning of the 1970s, special piteraq warnings have been issued by the Danish Meteorological Institute.

Surface air temperature in Greenland during summer is dominated by radiation effects as the sunlight returns to the Arctic. As a result, the mean temperature in July in the northernmost part of the country is only about 2 °C colder than in the southernmost part. However, temperatures at the coast are strongly influenced by the ocean and sea-ice variability; inland temperatures in ice-free regions can, therefore, be 5 °C warmer than at the coast. The proximity of the ice sheet does not seem to cool neighboring areas because the air flow off the ice during summer is usually warmer than the low lying local air mass due to adiabatic warming.

The latitudinal temperature gradient is much greater during winter than during summer with average sea level February air temperatures in the north of -36 °C and in the south of -4 °C. Variability is also much greater without the moderating influences of the open ocean in the north and the melting ice sheet. Katabatic storms often raise winter temperatures above freezing via the combined effects of adiabatic warming and

mixing of the inversion layer, particularly in the southern regions, where winter temperatures in the fjords can reach 10 °C (Cappelen, 2003).

Analysis of the long-term temperature record (1840 to 2001) from coastal stations in Greenland (Vinther et al., 2006, 2010) puts the more recent observations from satellites and weather stations on the Greenland Ice Sheet into a broader context (Box, 2002). The warmest decades in the long-term Greenland temperature record (1840 to 2000) are the 1930s and 1940s. Generally, the west coast experienced warming from 1873 to 1930 followed by cooling until 1988. There has been a warming trend since 1985 of 2 to 4 °C in West Greenland, primarily driven by winter temperature anomalies. The current warm period, beginning in 1988, is not unprecedented because the record warm decades occurred during the 1930s and 1940s.

The recent warmth occurred in spite of a persistently positive North Atlantic Oscillation (NAO) since around 1970. The NAO has a significant influence on the coastal temperature record particularly in the western and southern regions of Greenland during winter. The spectra of the temperature records display peak power at 3.7 years which Box (2002) attributed to the NAO, which has a spectral peak at ~4.0 years. The positive trend in coastal temperatures in southwestern Greenland appears to be responsible for the ice-sheet thinning observed in the region (Ohmura et al., 1999).

Volcanism has a significant cooling effect on temperatures in southern and western Greenland with a peak lag of 5 to 10 months following major eruptions (Box, 2002). The warming trend in Greenland since the early 1990s, in spite of major volcanic eruptions in 1982 and 1991, stands in contrast to the warmth of the 1930s and 1940s during a period of anomalously low volcanic activity (Overland et al., 2004).

Meteorological station records and regional climate model output were combined to develop a continuous 168-year (1840 to 2007) spatial reconstruction of monthly, seasonal, and annual mean Greenland Ice Sheet near-surface air temperatures (Box et al., 2009). Box and co-workers found that volcanic cooling episodes were concentrated in winter and along the western ice sheet slope, that inter-decadal warming trends coincided with an absence of major volcanic eruptions, and that 2003 was the only year in the 168-year series with a warm anomaly that exceeded three standard deviations from the 1951–1980 base period. The magnitude of the annual whole ice sheet 1919–1932 warming trend is 33% greater than the 1994–2007 warming. Box et al. (2009) also found that the recent warming was stronger along western Greenland in autumn and southern Greenland in winter. Spring trends marked the 1920s warming onset while autumn leads the 1994–2007 warming.

The climate on the Greenland Ice Sheet has been studied intensively in an effort to quantify the surface mass balance and to estimate its contribution to global sea level. However, in situ observations are limited due to the large area and remoteness of the ice sheet. The notable exception is the climate record from the Greenland Climate Network (GC-Net) of automatic weather stations, operating since 1995 and distributed across the ice sheet (Figure 8.4). Scientists are, therefore, forced to rely on remotely sensed observations and estimations from downscaled general circulation models (GCMs) or high-resolution regional climate models (RCMs). However, systematic biases in the models relative to the in situ record illustrate that the climate of the ice sheet is more or less decoupled from the coastal regions where most of the direct observations that feed the models are made. Nevertheless, a consensus picture of change due to a warming climate since 1995 is emerging.

8.2.1.2. Climate observations

The Danish Meteorological Institute (DMI) data are the most comprehensive meteorological records available for Greenland's coastal region and have been homogenized, most notably the original observations have been checked and the data compared with time series of related climatic elements for the same stations, with the specific purpose of studying long-term climatological trends (Cappelen et al., 2001).

Monthly air temperature records from eight DMI synoptic stations at the coast of Greenland (located mainly in the southwest but including Tasiilaq in the southeast) show a pronounced warming since the early 1990s. The warming follows an overall trend of regional cooling between 1958 and 1992, concentrated in the winters of the 1960s to 1980s (Hanna and Cappelen, 2003). Summer trends for the 7-station DMI average

(i.e., excluding Kangerlussuaq because of its relatively short record) indicate a significant warming for 1961 to 2006 of 0.9 °C.

For the DMI station average, 2003 had the warmest summer (i.e., June to August) on record, with a mean temperature of 8.1 °C at 3.3 standard deviations above the most recent climatological 'normal' period (1971–2000) mean. The second warmest summer occurred in 2005 (7.6 °C) at 2.5 standard deviations above the 1971–2000 mean. 2005 was more than 0.5 °C warmer than the third warmest summer, 2006 (7.07 °C) which was closely followed by 2001 (7.02 °C), 1965 (7.00 °C), and 2004 (6.97 °C) (Hanna et al., 2008).

The record for the GC-Net automatic weather station 'Swiss Camp' (1170 m above sea level in 1991 and since then decreasing 0.32 m annually; Stober and Hepperle, 2006; Figure 8.4) (Steffen and Box, 2001) was used to gauge temperature changes on the western flank of the Greenland Ice Sheet, where extensive seasonal melt and relatively high runoff from this relatively low elevation zone contribute a large proportion of the total runoff (Hanna et al., 2005; Box et al., 2006). This record, by far the longest GC-Net series, spans 19 years (1991 to 2009), and its interannual variability is significantly correlated with that of the mean of the seven DMI coastal stations (de-trended series r = 0.65, $p < 0.01$; Figure 8.5).

Similar to the positive (7-station mean) DMI temperature trend, Swiss Camp summer mean temperature increased significantly by 2.4 °C from 1991 to 2008, or 2.2 °C for the period 1993 to 2008, excluding the global cooling effect of the Mount Pinatubo eruption in 1992. The three summers of 2003/2004/2005 were almost equally record warm (mean temperatures 0.3, 0.3, and 0.2 °C, respectively) at Swiss Camp, alongside 1995 (0.5 °C) and the record year 2007 (1.2 °C). The latter season has been previously noted for its relatively high modeled runoff compared with most other years between 1958 and 2003 (Hanna et al., 2005). The latest data from the Swiss Camp GC-Net record show a continued warming with another record mean summer temperature of 1.2 °C in 2007; 0.7 °C above the previous maximum in 1995.

A reprocessed and updated near-surface air temperature series for Summit (1987–2005, 3205 m above sea level) (Shuman et al., 2001) shows a slight overall -0.5 °C cooling in summer, in contrast to all the other Greenland temperature records (all from much lower elevations and generally around the ice sheet margin; Figure 8.5). This new Summit series is a re-analyzed composite, primarily of University of Wisconsin and ongoing GC-Net automatic weather station data, supported by SSM/I (Special Sensor Microwave / Imager) brightness temperature data (Shuman et al., 1996, 2001). There is a highly significant correlation between individual years' fluctuations in the de-trended DMI and the Summit series (Figure 8.5).

Three hypothesized possible causes for the cooling trend observed at Summit compared to the warming trend along the coast are: (i) continued relative suppression of more regional climatic change by thermal inertia of the huge central Greenland ice mass as noted for Arctic Ocean sea ice (Serreze and Francis, 2006); (ii) a differential response of the high elevation zones of the Greenland Ice Sheet in accordance with the well-known lower tropospheric warming / higher atmospheric cooling response to increased greenhouse gases (Stott et al., 2006), which has been demonstrated specifically for Greenland in a recent analysis of radiosonde data spanning the period 1964 to 2006 (Box and Cohen, 2006); and/or (iii) regional changes in wind, cloud cover, or radiation patterns over the Greenland Ice Sheet.

This observed pattern (coastal, i.e., marginal, warming combined with little change or slight cooling in the high interior) is opposite to the output of simulations from Atmosphere-Ocean General Circulation Models (AOGCMs). In all high-resolution temperature changes studied in the present century, Huybrechts et al., (2004) and Gregory and Huybrechts (2006) found that modeled summer warming is actually greatest over the interior of Greenland and least along the coast.

High-resolution (5 km × 5 km) surface air temperature (SAT) data were bilinearly interpolated from the 0.5° resolution 40-year European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and corrected for ECMWF terrain errors using empirically derived ice sheet surface lapse rates, as explained by Hanna et al. (2005). These SAT data provide corroborating evidence for anomalously (3 °C) high summer temperatures around the Greenland margins during recent warm summers with a noted high melt, which occurred most widely within the southern and western marginal ablation zones of the ice sheet.

The accumulation zone of the Greenland Ice Sheet (above 2000 m elevation) apparently had cold summer anomalies in 2003 and 2006. This is in contrast to warm anomalies in more outlying areas but in line with the Summit temperature results discussed previously. Summer temperatures both in Greenland and across the Northern Hemisphere exhibit strongly rising trends since the early 1990s, although the earliest part of this period (1992–1993) characterizes the general temperature recovery following the cooling after the 1991 Mount Pinatubo volcanic eruption (Robock and Mao, 1995).

Observations and models both indicate the occurrence of recent high snow accumulation events in the winter of 2004/2005, concentrated in western Greenland (Nghiem et al., 2007), and winter-spring 2002/2003 in southeastern Greenland (Krabill et al., 2004; Hanna et al., 2006). Huybrechts et al. (2004) hypothesized that such events may become more frequent in Greenland as storm tracks intensify or shift position with climate change.

8.2.1.3. Surface and basal properties and processes

Surface melting in the ablation zone (Figure 8.6) is highly sensitive to changes in surface albedo. Melting of ice in the ablation zone exposes a surface layer of dust that was originally deposited with snowfall high on the ice sheet. Dust in snow and on ice reduces the solar reflection (albedo) and, thus, increases surface melt further. Surface albedo decreased dramatically at Swiss Camp during 1993 to 2008 and this led to increased surface melt and ablation at Swiss Camp (Figure 8.7).

Basal melt and meltwater exert a strong influence on ice flow (Fahnestock et al., 2001). The limited knowledge of basal melt is derived from models and sparse observations of bed properties. A basal melt rate of 0.1 m/y over a substantial area requires a regional geothermal heat flux of 970 MW/m^2, which is 17 times higher than the continental background of 56.5 MW/m^2 (Sclater et al., 1980). This does not account for heat conducted through the ice. For most of the ice sheet, the heat conducted through the ice is close to the background geothermal heat flux. It can be lower in low-accumulation regions because the thermal gradient in the basal ice is lower. The volume of meltwater being produced in the areas of high basal melt is of the order of 1 km^3/y. This water and warm basal ice are responsible for the onset of rapid ice flow in the large ice stream that drains the north side of the summit dome in Greenland (Fahnestock et al., 2001). Strong melt also occurs due to deformation heat under the fast-flowing ice streams thereby further increasing their flow. Melt rates of metres per day are observed where the base of the ice streams is in contact with the ocean (Rignot et al., 2010).

Basal flow can transport ice at velocities exceeding rates of internal deformation (i.e., from hundreds to more than 10 000 m annually) and glacier surges, tidewater glacier flow, and ice stream motion are governed by basal flow dynamics (Clarke, 1987). Glaciers and ice sheets that are susceptible to basal flow can move quickly and erratically, making them intrinsically less predictable than those governed by internal deformation. They are more sensitive to climate change because of their high rates of ice turnover which gives them a shorter response time to climate (or ice margin) perturbations. In addition, they may be directly responsive to increased amounts of surface meltwater production associated with climate warming. This latter process is crucial for predicting dynamic feedbacks to the expanding ablation area, longer melt season, and higher rates of surface meltwater production that are predicted for most ice masses.

Although basal meltwater is most likely to be the primary source of subglacial water, models have shown that supraglacial streams with discharges of over 0.15 m^3/s can penetrate down through 300 m of ice to reach bedrock, via self-propagation of water-filled crevasses (Arnold and Sharp, 2002). There are several possible subglacial hydrological configurations: ice-walled conduits, bedrock conduits, water film, linked cavities, soft sediment channels, porous sediment sheets, and ordinary aquifers (Mair et al., 2001; Flowers and Clarke, 2002).

Modern interest in water flow through glaciers originated in a pair of theoretical papers published in 1972. In one, Shreve (1972) discussed the influence of ice pressure on the direction of water flow through and under glaciers, while Röthlisberger (1972) presented a theoretical model for calculating water pressures in subglacial conduits in the other. Through a combination of these theoretical considerations and field observations, Röthlisberger concluded that the englacial drainage system probably comprises an arborescent

network of passages. The millimetre-sized fingertip tributaries of this network join downward into ever larger conduits.

Locally, moulins provide large direct connections between the glacier surface and the bed. Beneath a valley glacier, the subglacial drainage is likely to occur as a tortuous system of linked cavities transected by a few relatively large and comparatively straight conduits. The average flow direction in the combined system is controlled by a combination of ice overburden pressure and bed topography, and, generally, is not normal to contours of equal elevation on the bed.

Luthje et al. (2006) studied the summer evolution of supraglacial lakes on the Greenland ice margin using a one-dimensional model to calculate the surface ablation for a bare ice surface and beneath supraglacial lakes for 30 days in the summers of 1999 and 2001. The surface ablation beneath the lake was enhanced by 110–170% for the two years 1999 and 2001 compared with the ablation for bare ice. Within the region of the ice sheet where supraglacial lakes presently occur, the area covered by supraglacial lakes was found to be more or less independent of the summer melt rate but controlled by topography. Luthje and co-workers predicted that, inside the ablation region, the area covered by supraglacial lakes will remain constant even in a warmer climate.

Although theoretical studies usually assume that subglacial conduits are semicircular in cross section, there are reasons to believe that this ideal is rarely realized in nature. Much of the progress in subglacial hydrology has been theoretical, as experimental techniques for studying the englacial hydraulic system are few and, as yet, not fully exploited and observational evidence is difficult to obtain. How directly and permanently do these effects influence ice dynamics? It is not clear at this time. The process is well known in valley glaciers where surface meltwater that reaches the bed in the summer melt season induces seasonal or episodic speed-ups (Iken and Bindschadler, 1986). Speed-ups have also been observed in response to large rainfall events (e.g., O'Neel et al., 2005).

8.2.1.4. Summary

Is the climate in Greenland changing? Coastal records since 1840 show the warmest decades to be the 1930s and 1940s and a warming trend since 1988 of 2 to 4 °C in West Greenland, mainly during the winter months. Recent, short-term temperature records on the ice sheet in West Greenland (Swiss Camp) show a 2.2 °C warming in summer since 1991 with the highest air temperatures in 2007. Air temperatures at the top of the Greenland Ice Sheet (Summit) revealed a slight cooling of -0.3 °C between 1987 and 2005. Decadal-scale trends over the past century, however, are complicated by the large degree of natural variability (Chylek et al, 2006). The observation period for the recent climate warming over Greenland is short.

8.2.2. Surface mass balance

- The surface mass balance of the Greenland Ice Sheet has decreased since 1990. The mean of three model reconstructions of the surface mass balance show a 40% decrease from 350 Gt/y (1970–2000) to 200 Gt/y in 2007.

- The runoff has increased significantly due to higher temperatures, increasing ablation zone area and a longer melt season.

- These trends in surface mass balance have been linked to a global warming trend that exceeds the local natural variability in climate.

8.2.2.1. Introduction

The total mass balance of the Greenland Ice Sheet (Section 8.2.4) determines its overall state of health and is the sum of two terms: the surface mass balance (SMB), discussed here, and solid ice discharge from marine terminating glaciers, discussed in Section 8.2.3. The SMB is the net mass added or removed from the surface of an ice sheet by a range of processes. The aim of this section is to introduce the factors that control SMB, and how they respond to climate change. This is achieved by reviewing recent estimates of the components

that make up the SMB, discussing the degree of consistency in these components, both from observations and modeling, and considering areas of uncertainty requiring further development.

The annual SMB is defined as the sum of mass fluxes to and from the ice sheet surface, integrated over a year and over the area of the ice sheet. Mathematically, this is expressed as:

$$SMB = P + TMT - R - Q$$

where P is total precipitation, comprising the solid fraction (snow, hail, freezing rain) as well as the liquid fraction (rain); TMT is turbulent moisture transport, representing the net effect of surface evaporation, sublimation/deposition and snowdrift sublimation; R is runoff, representing the net effect of condensation, melt, refreezing and retention; and Q is the snowdrift blown off the ice sheet integrated along the entire margin. All fluxes are in Gt/y and integrated over the area of the ice sheet. Snowdrift plumes being blown into the ocean have been occasionally observed, but Q is believed to impact SMB significantly only locally and is not further considered here (Box et al., 2006). The accumulation zone is the region where SMB > 0 and the ablation zone is the region where SMB < 0. The equilibrium-line altitude (ELA) is the elevation where SMB = 0. It is important to note, however, that precipitation takes place in the ablation zone and runoff can take place in the lower part of the accumulation zone. Above the ELA, however, accumulation (via snowfall) exceeds ablation (via runoff) and below the ELA the opposite is true. For an ice sheet as a whole to be in a state of balance, SMB must match the solid ice flux crossing the grounding line and flowing into the ocean. For the Greenland Ice Sheet, this was roughly the case prior to the 1990s but there has since been a marked increase both in solid ice flux (Rignot and Kanagaratnam, 2006) and runoff (e.g., Hanna et al., 2008; Ettema et al., 2009). Accumulation includes all processes by which mass is added to the ice sheet, which, in this case is almost entirely through precipitation in the form of snowfall. The accumulation rate is what is typically measured with in situ observations from snow pits or shallow ice cores in the high elevation zones of the ice sheet where there is basically no runoff. It is usually expressed as a water equivalent (w.e.) thickness change per unit area per unit time. Ablation includes all process by which mass is lost from the ice sheet. This section, however, only considers ablation via surface melting (called runoff). The other major component of ablation due to solid ice discharge into the ocean is dealt with in Section 8.2.3. The ablation rate is identical and has the same units but is a measure of net mass loss at the surface rather than mass gain. Ablation rates can be measured during the summer melt season using, for example, stakes drilled into the ice surface. These observations of surface accumulation/ablation rate are equivalent to the local SMB.

There are a number of approaches for estimating SMB. The first, and the only tractable approach until quite recently, involves the interpolation of in situ observations of point SMB (accumulation/ablation) estimates from snowpits, ice cores, stake measurements and automatic weather stations to estimate accumulation rates (Figure 8.8). The data were combined with simplified models of runoff, R, to determine SMB (Ohmura and Reeh, 1991; Reeh, 1991; Ohmura et al., 1999; Bales et al., 2001; Steffen and Box, 2001). The more recent interpolations to determine accumulation rates also included a formal assessment of the error associated with field data and their interpolation (Cogley, 2004; Bales et al., 2009) as discussed in Section 8.2.2.2.

Commonly, runoff has been estimated using an empirical model of melt, M (see Section 8.2.2.3), combined with simplified assumptions about refreezing, RF, calibrated using stake measurements (Reeh, 1991; Braithwaite et al., 1992; Braithwaite, 1995). The differences in the estimations of runoff from the different models are of the order 25% (Table 8.2). The individual components contributing to runoff (M and RF) vary by far more than this. Thus, the agreement in R is largely based on compensating differences. The different runoff models have very different sensitivities to forcing and, as a consequence, the ability to predict the true response of SMB to a future change in climate also contains significant uncertainties (Bougamont et al., 2007).

Table 8.2. Estimates of the SMB components in Gt/y for Greenland from satellite (Mote, 2003) and modeling studies for the range of time periods specified. The regional climate models, PMM5, RACMO2/GR and MAR are discussed in Section 8.2.2.2. The sum of RF and R does not necessarily equal M, depending on how rain is treated.

Model	Period	Area × 10^7 km^2	P	TMT	Rain	M	R	RF	SMB

MAR[a]	1958–2007	1.701	575	-5	20	532	282	250	264
PMM5[b]	1958–2006	1.691	555	-63	16	228	213	48	326
Hanna[c]	1958–2007	1.678	559	-35	26	313	261	77	264
RACMO2/GR[d]	1958–2007	1.69	743	-26	46	404	248	199	469
Mote[e]	1988–1999	1.648	591	-69	23		255		239
Reeh[f]	1990s	1.707	552[g]				279		273

P: Precipitation, comprising the solid fraction (snow, hail, freezing rain) as well as the liquid fraction (rain); TMT: Turbulent Moisture Transport, comprising surface evaporation, sublimation/deposition and snowdrift sublimation; Rain: liquid fraction of precipitation; M: total volume of Meltwater; R: Runoff, comprising the liquid water fraction from melt and rain that leaves the firn/ice column; RF: Re-Freezing; SMB: Surface Mass Balance.

[a] Modèle Atmosphérique Régional (Fettweis, 2007); [b] Polar Mesoscale Model version 5 (Box et al., 2004); [c] Hanna et al. (2002, 2005); [d] Ettema et al. (2009); [e] Mote (2003); [f] Reeh et al. (1999); [g] Reeh and co-workers (1999) did not estimate the individual terms in accumulation and the tabulated value represents, therefore, P-TMT-Rain.

Recently, alternative approaches for estimating SMB have been developed, especially following the release of a high-quality, consistent ~50 year global climatology produced by ECMWF. The data set is known as ERA-40 (i.e., ECMWF Re-Analysis covering the ~40 year period 1957-2002). ERA-40 is a hind-cast generated using the ECMWF atmospheric forecast model driven by surface and satellite meteorological observations (Uppala et al., 2005). ERA-40 produces 6-hourly and daily fields of key atmospheric parameters at 1.125° resolution, which equates to a grid cell size of ~125 km in latitude. This resolution is too coarse to be used unadjusted for SMB calculation as the ablation zone, for example, is mostly less than a single grid cell in width. As a result, various approaches, commonly termed downscaling, have been used to interpolate ERA-40 to higher resolution. The simplest approach has been to downscale surface air temperature using seasonally adjusted lapse rates and high-resolution surface topography as well as a simple interpolation of precipitation (Hanna et al., 2002, 2005) (Table 8.2 and Figure 8.9).

Another approach is to use an atmospheric RCM as a physically-based 'interpolator' of the reanalysis data. Generally, an RCM is run at high resolution for a limited area and forced at its boundaries by climate data from a coarser-resolution GCM. To date, three RCMs have been used to produce a high-resolution time series of the Greenland Ice Sheet SMB (Box et al., 2004, 2006; Fettweis, 2007; Ettema et al., 2009) although others are under development.

It is interesting to note how SMB responds to changes in climate. Increased near-surface air temperatures bring about increased moisture and, therefore, precipitation. Simulations with coupled atmosphere-ocean GCMs suggest a 5% /K sensitivity of P to surface air temperature (Gregory and Huybrechts, 2006). This value can be combined with SMB 'climate sensitivities' (especially to changes in temperature and precipitation) estimated from an energy balance model that suggests that the area-averaged SMB sensitivity, CT, to temperature is -49 mm/K and to precipitation, CP, 3.8 mm/% (Oerlemans et al., 2005). Using the sensitivity of precipitation to temperature, given above, makes it possible to express CP with respect to temperature, which gives CP = 19 mm/K. Thus the increase in runoff greatly outweighs the increase in accumulation (49 *vs* 19) for a fixed temperature change. Assuming the area of the ice sheet to be 1.7 million km^2 then the increased mass loss from the ice sheet is 47 Gt/K, excluding any dynamic response. Thus, simulations of future SMB for various greenhouse gas emissions scenarios generally predict a decreasing SMB (Gregory and Huybrechts, 2006). Most importantly, this is what has been observed to have taken place over the past 15 years, since the temperature increase over Greenland started (Hanna et al., 2008).

8.2.2.2. Accumulation

8.2.2.2.1. Observations

Shallow ice cores and snowpit measurements have been collected and used to determine accumulation rates, going back to the 1920s (Ohmura and Reeh, 1991). In this compilation, 251 measurements of accumulation rate were used to derive a map of precipitation for the whole of Greenland. Since the 1990s, the Program for Arctic Regional Climate Assessment (PARCA), a NASA-funded, multi-decadal, coordinated project, has been supplementing these records and collecting many other types of data with the aim of improving estimates of a wide range of key parameters for the ice sheet (Abdalati, 2001). Among the measurements

taken were over 100 shallow ice core estimates of accumulation. These were combined with a subset of the historical records to produce the most comprehensive observation-based map of accumulation across the ice sheet to date. The estimated mean error for this compilation was 24% with regional variations of 15–30% (Bales et al., 2001). This compilation has recently been updated with a number of new cores and coastal station data (Figure 8.10) (Bales et al., 2009). The recent update resulted in regional changes, compared with the earlier study, of as much as ±50%, suggesting that uncertainties in accumulation rates were significantly underestimated in previous studies: a conclusion that is corroborated by the substantial differences in modeled accumulation shown in Figure 8.9.

The average accumulation rate of 300 kg/m^2 over the ice sheet agrees well with an earlier estimate obtained from different sources and covering a different time period (Ohmura and Reeh, 1991), and with a further estimate using a different interpolation approach (Cogley, 2004). Although there is agreement between the average values of the estimated accumulation, examination of the spatial distribution demonstrates marked and significant differences (Figure 8.8).

An error analysis of in situ snowpit and ice core data suggested an uncertainty of 7.7% in the averaged 30-year mean. This is in agreement with the error estimate from the most recent study (Bales et al., 2009). It does not, however, reflect the interannual and spatial uncertainties which are very much greater (Bales et al., 2009). Uncertainties in the spatial pattern can have a significant impact on modeled estimates of ablation due to the influence of snowfall on surface albedo and refreezing (see discussion in Section 8.2.2.3). In addition, rain is a small but important term of the accumulation rate. Although its volume is low (~4% of total precipitation), its impact can be important because upon refreezing it raises the heat content of the snowpack, and in doing so increases total melt and runoff volume. The increased percentage of precipitation falling as rain in a warming climate will have a stronger influence on runoff and, hence, SMB.

8.2.2.2.2. Regional climate models

One big advance in assessing the SMB of the Greenland Ice Sheet, its temporal evolution and the relative importance of different components and their sensitivity to climate has been the development of RCMs. The first high-resolution RCM for Greenland was based on a modified version of the fifth generation mesoscale model known as PMM5 (Polar MM5; Box et al., 2004). This model ran at 24 km resolution and was calibrated using in situ data to correct for biases in melt energy and water vapor fluxes (Box et al., 2004). It was run, initially, for the period 1957 to 2004 forced by ERA-40 data (Box et al., 2006) (see Table 8.2 and Figure 8.9).

More recently, a 25 km resolution RCM was developed by Gallee and Schayes (1994) and applied to the Greenland Ice Sheet by Fettweis (2007) for the period 1958 to 2006. Unlike PMM5, the MAR (Modèle Atmosphérique Régional) has not been calibrated or corrected using in situ data. Comparison of modeled water vapor fluxes from MAR with automatic weather station data suggests that TMT is underestimated in MAR (Fettweis, 2007). Precipitation and runoff, however, are broadly comparable with other estimates from modeling and observations. The individual terms that comprise the runoff (M and RF) are both much greater than, for example, those of Hanna et al. (2008) (Table 8.2, Figure 8.9).

The third, a 50-year RCM reconstruction driven by ERA-40 data, is known as RACMO-2/GR. This is an adapted version of the ECMWF regional forecast model and was initially developed to run over Antarctica (van de Berg et al., 2006). More recently it has been coupled to a snow metamorphism and melt model (Bougamont et al., 2005) and run at 11 km resolution over a domain covering the whole of Greenland (Ettema et al., 2009). Like MAR, it was not calibrated using any in situ observations. The total accumulation from the RACMO-2/GR RCM is 7–24%, higher than the estimates from the other RCMs (Figure 8.9c). The greatest differences between the accumulation estimates from the RCMs are found in the wet southeast of Greenland where precipitation is estimated to exceed 4 m w.e /y locally. This unusually high value is consistent with a single ice core obtained from this region (Ettema et al., 2009).

It should be noted that RCMs are the only approach where all the components of SMB can be separated. In situ data provide the net SMB: that is, precipitation minus TMT terms and runoff. Also, measuring melt volume and refreezing, in the field, is challenging (Bøggild et al., 2005). The models suggest that TMT is

generally negative, indicating net sublimation / evaporation, with a magnitude of about 10% of P, except in MAR, where it is an order of magnitude smaller.

8.2.2.3. Ablation

8.2.2.3.1. Observations

Direct observations of ablation rates around the margins of the Greenland Ice Sheet are sparse both in space and time (Figure 8.11). Ablation has been measured using stakes at a number of sites in Greenland (Braithwaite et al., 1992) and, more recently, a technique has been developed for making automated measurements (Bøggild et al., 2004a). However, the spatial coverage is far from representative. One of the more comprehensive and uninterrupted in situ data sets has been collected since 1990 by Utrecht University along a 150-km transect near Kangerlussuaq, central western Greenland, known as the K-transect (Greuell et al., 2001; van de Wal et al., 2005). Stake measurements of mass balance have been collected at eight sites along the K-transect since 1990 and automatic weather station data are available since 1997 at one site and since 2003 at two further sites.

This represents the longest existing continuous record of in situ mass balance and ablation data for the Greenland Ice Sheet, but how representative the data are of other parts of the ablation zone is uncertain. This region of the ablation zone is, generally, smoother and less rugged compared with areas to the north and south. Small-scale topography affects the microclimate and therefore local ablation variability. A modeled time series used to assess the spatial autocorrelation of SMB showed little autocorrelation in the ablation zone in the southeast, a negative correlation in the northeast, and a strong positive correlation along the western margins suggesting that ablation depends on local conditions (as would be expected) as well as larger-scale climate forcing (Box et al., 2006). For example, in northeastern Greenland at the Storstrømmen glacier at 77° N, the observed large spatial variability in ablation is mainly controlled by the katabatic windfield producing a highly uneven winter snow cover distribution and higher melt rates locally at convex surface undulations.

Although the in situ data cannot be used to provide direct estimates of ice-sheet integrated ablation or the individual ablation components, they can be used to validate and calibrate satellite and model-based estimates. The spatial coverage of the quite sparse observational points, however, is currently not adequate to provide representative sampling to validate and calibrate satellite and model-based estimates of ablation.

The runoff component of the ablation has also been estimated using calibrated satellite observations of passive microwave melt extent and duration (Mote, 2003) and satellite-derived albedo data (Greuell and Oerlemans, 2005). Both approaches rely on an empirically derived relationship between melt duration or albedo and runoff. Estimating runoff from routinely acquired satellite observations is a promising technique but is empirically based and it remains to be shown whether the relationships derived are robust under current and future (changing) climate conditions. Recent results have shown, for example, that for areas in Antarctica where melt occurs, passive microwave-derived values underestimate SMB (Magand et al., 2008).

In addition to estimating runoff, passive microwave data have also been used to measure the area of the Greenland Ice Sheet subject to surface melt (Abdalati and Steffen, 2001; Mote, 2007). Results indicate that the melt area has been steadily increasing since satellite observations began (Figure 8.12). Although this is not a direct measure of runoff (because it does not estimate the volume of melt or refreezing) there is a good correlation between melt area, number of melt days, and runoff (Mote, 2003).

In this approach, the satellite-based melt duration data are used as a proxy for surface air temperature. Although this method has some merit, the development of a consistent, homogeneous temperature data set derived from ERA-40 and extending back to 1958, provides a longer time series with roughly the same correlation with runoff (Hanna et al., 2002, 2005). Its advantage, however, is that it represents a direct observation of changing surface properties as opposed to a model reconstruction and can, therefore, be used to corroborate the estimates derived from re-analysis data. In addition, it provides observational evidence for changes in the length of the melt season and its onset date (Figure 8.12).

8.2.2.3.2. Runoff modeling

The first reasonable, quantitative attempts to model runoff from the Greenland Ice Sheet used a positive degree day (PDD) approach (Reeh, 1991). Here, the melt volume is related to the number of days where the near surface air temperature (usually taken at 2 m above the surface) exceeds 0 °C, and by how much. Melt volume is calculated by multiplying the positive temperature by a degree day factor (typically ~ 8 mm/°C per day for ice). This approach has been, and still is, used extensively for modeling the present-day runoff and SMB of the Greenland Ice Sheet (Braithwaite, 1995; Abdalati et al., 2001; Mote, 2003; Hanna et al., 2006; Rignot and Kanagaratnam, 2006) as well as its response in the future (Huybrechts et al., 2002, 2004).

The compelling advantage of the PDD approach is that it relies on a single, smoothly varying and easily interpolated parameter, i.e., near surface air temperature. Climatological maps of this parameter have been produced from a combination of coastal station and inland automatic weather station data (e.g., Ohmura, 1987) and more recently downscaled from ERA-40 (Hanna et al., 2002, 2005). The weakness of the approach is that it relies on an empirically based relationship, calibrated using a limited range of data for present-day conditions only. As a result, it is unclear how well a PDD model can predict melt in the future (Bougamont et al., 2007). For example, a PDD model has no sensitivity to changes in cloud cover or type unless they impact near-surface air temperature. Modifications to the approach have been suggested to incorporate more of the physical processes involved (Pellicciotti et al., 2005) but remain, fundamentally, empirical.

A more physically rigorous approach is to calculate the energy balance at the surface and, from this, the quantity of energy available for melting. Most recently, energy balance models have been used to estimate melt volume over Greenland by downscaling meteorological hindcast data such as ERA-40 (Box et al., 2004, 2006; Fettweis et al., 2005). Generally, these types of model incorporate all the key processes that affect melt but, as a result, require a large number of meteorological inputs that may not be constrained particularly well. For example, to estimate the turbulent, sensible and latent heat fluxes, requires prescribing a surface roughness as well as atmospheric stability corrections. It also requires near-surface humidity and the vertical wind profile in the boundary layer, all of which may not be particularly well known.

The single most important surface property for controlling runoff in the energy balance, however, is surface albedo, α. This can be estimated by the model based on snow age and/or density (Greuell and Konzelmann, 1994; Konzelmann et al., 1994) or prescribed from satellite observations (Stroeve, 2001). Model-based estimates, however, are generally only reliable for snow. Ice in the ablation zone can have an albedo of 0.4 or less (see Figure 8.7), which is substantially less than that of snow. Ice albedo depends on aerosol content and this is highly spatially heterogeneous. The advantage of an energy balance model compared to a PDD approach is that it incorporates the physical processes responsible for melt explicitly and should, therefore, respond appropriately to changing climate forcing. To do this, however, requires the 'correct' forcing. For long integrations (e.g., multi-centennial) such input data typically do not exist or may not be reliable enough to justify the use of a more complex and more 'data hungry' energy balance model. Thus, although an energy balance model may provide a more physical representation of the processes, it will not always be the best tool to use.

Preconditioning of the snowpack and the firn is very important for subsequent melt and runoff. In the SMB time series of Hanna et al. (2005), high runoff years (except 2003) are generally synchronous with low precipitation / accumulation and vice versa. More accumulation results in a higher albedo for a longer time, which, in turn, reduces absorbed energy available for melt; the available surface energy needs to melt any snow first before ice can melt in the ablation region. Also, higher volumes of meltwater are retained in the thicker snowpack, which tends to reduce net runoff. Thus, accurate reconstruction of the timing, pattern and volume of accumulation is important for reliable estimation of runoff.

Another important consideration in modeling runoff is refreezing of meltwater within the snowpack. A variety of approaches exist from the simplest, using a fixed fraction of the annual accumulation (Reeh, 1991) to full snow metamorphism models (Bougamont et al., 2005; Mernild et al., 2006; Fettweis, 2007). The latter contain explicit terms for the metamorphism of the snowpack as meltwater percolates into it but can, in general, be driven only by an energy balance model because they require the energy balance of the snowpack to be calculated, which is not possible using a simpler PDD approach. A study of three relatively simple refreezing schemes for the Greenland Ice Sheet found that they accounted for about 20% of the total melt

production, produced similar estimates, and had reasonably similar sensitivities to changing climate (Janssens and Huybrechts, 2000).

More recent analyses incorporating these simple approaches as well as snow metamorphism models reached a different conclusion, however, and found that the different models produced different volumes of refreezing and had different sensitivities (Bougamont et al., 2007; Wright et al., 2007). The MAR RCM, for instance, estimates that refreezing is about 50% of the total melt production (Table 8.2). These studies suggest that further work on understanding and modeling refreezing is required. They also indicate that modeling refreezing remains a significant uncertainty in present-day and future estimates of SMB. A major hurdle for improving refreezing estimates is both the difficulty in obtaining reliable in situ validation data (Bøggild, 2000) and the difficulty of incorporating preferential water flow deep into the snowpack with the models, which is known to take place (Bøggild, 2000). Instead of refreezing in the cold snow, water channels into vertical flow fingers followed by horizontal conduits and eventually results in runoff (Bøggild, 2000).

8.2.2.4. Differences and trends in measured and modeled SMB

The average SMB for the four reconstructions (RACMO2/GR, MAR, PMM5, Hanna; see Table 8.2) covering the past 50 years is 354 Gt/y with a range of 206 Gt between the lowest and highest estimates. This range is 58% of the mean value and provides some measure of the uncertainty in the SMB. This compares with a mean estimate of precipitation of 599 Gt/y. Thus, roughly half the precipitation is lost as runoff during this period, although estimates vary substantially (Table 8.2). It should be noted that runoff is not equal to the total volume of meltwater, M, because a proportion of this percolates into the underlying snowpack and refreezes. Considerable uncertainty exists in estimating this refreezing term, RF (see Section 8.2.2.4).

Table 8.2 summarizes some recent estimates of Greenland SMB, derived from a variety of methods. It is important to note that the values presented in Table 8.2 are not entirely independent. The models labeled Mote and PMM5 derived accumulation rates from almost identical downscaling approaches. Hanna, MAR, RACMO2/GR and PMM5 all use the same source: ECMWF climate re-analysis (known as ERA-40) and operational analysis data as the input for their downscaling approach. Thus, biases in ERA-40 may affect all of these estimates in similar ways.

Estimates of SMB from downscaling of ERA-40 data agree reasonably well when averaged over several years but there are substantial differences in spatial patterns, individual components of SMB, and interannual variability. Figure 8.9 compares the components of SMB from several different approaches: A simple downscaling of ERA-40 derived by Hanna et al. (2005) and the RCMs, MAR (Fettweis, 2007), RACMO2/GR (Ettema et al., 2009) and PMM5 (Box et al., 2006). It is not surprising that the trends are in agreement and that there is reasonable correlation between minima and maxima as this is largely constrained by the common forcing used in the simulations (i.e., ERA-40). There are, however, substantial departures in absolute values. For 2003, for example, the Fettweis (2007) SMB anomaly is about 180 Gt less than Hanna and PMM5, while RACM2/GR is consistently even higher due to its higher estimate of accumulation (Figure 8.9d). These differences are about as large as the SMB for that year.

The difference between the estimates is thus larger than the loss of mass from the ice sheet over the past decade (Section 8.2.5).

In addition, the refreezing component, RF, is highly uncertain. The range for RF between the four time series that use ERA-40 data as input is from 39 to 238 Gt/y: a difference that is 58% of the mean SMB. Refreezing is, therefore, an important, yet poorly known, component of SMB, which has received relatively little attention compared with other terms, partly due to the difficulty in making direct measurements of its magnitude. Although the fundamental physical processes that influence SMB are, in the main, relatively well understood, Table 8.2 and Figure 8.9 clearly show that much uncertainty still remains in constraining it.

Figure 8.13 summarizes the main trend in SMB over the past 50 years. Hanna et al. (2008) estimated a statistically significant positive trend in precipitation from 1958 to 2006 of 84 Gt/y, equivalent to 1.7 Gt/yr^2. Runoff, however, has increased at a greater rate, particularly since the early 1990s (Figure 8.9) resulting in a net reduction in SMB during this time (Figure 8.13) as expected in a warming climate. The modest increase in precipitation is supported by in situ measurements from an ice core in North East Greenland (Figure 8.14),

which show an increase since about 1990 once interannual variability is reduced by smoothing. However, observational data from one site is not enough owing to climate variability and significant noise in the data (Fisher, 1985; Andersen et al., 2006). There are, however, few studies extending to the most recent years and the data illustrate the need for many and long records before changes in accumulation can be statistically verified. The interannual variability in precipitation from the RCMs is around 20% and, as can be seen in Figure 8.9, means that short-term signals over, for example, the past decade are strongly influenced by this variability.

The SMB has a 50-year mean average value of 354 Gt/y. During the 1960s and early 1970s, SMB was 10–20% below the 50-year mean. Since the mid-1990s, however, SMB has decreased steadily to its lowest value over the entire reconstructed period, such that the most recent years are more than 100 Gt below the 50-year mean and 200 Gt below the value in the mid-1990s. This is the most striking trend in the time series and has been linked to a global, rather than regional, warming trend that exceeds the background natural variability (Hanna et al., 2008). This is an important conclusion, which suggests that the trend is likely to continue into the future. Also, crucially, the increased melt during this period has warmed the surviving snowpack at lower elevations by some 5 to 10 °C, priming it for even greater losses in the near future (van den Broeke et al., 2009).

8.2.2.5. Summary and recommendations

The mean SMB, based on the four ~50-year records, is 354 Gt/y with a range between estimates that is 54% of the mean, providing a qualitative indication of the uncertainty in the SMB. This is of the order of magnitude of the standard deviation of the interannual variability, which ranges from 62 Gt/y for PMM5 to 124 Gt/y for MAR. Since the mid-1990s, runoff has increased significantly with only a modest change in accumulation, resulting in a reduction in the SMB of around −200 Gt over the past 13 years (Figure 8.13). This trend in SMB appears to be a response to global climate change and above the level expected due to natural variability (Hanna et al., 2008). It should be noted, however, that the gradient of the trend is dependent on the epoch, due to the large interannual variability discussed above. Also of note, is that 2007 had the lowest SMB of any year in the 50-year time series. Significant differences exist between the various estimates of the components of the SMB derived from numerical modeling and downscaling of climate re-analysis data. These differences can be as large as the actual SMB estimate for a given year and the standard deviation of the differences is of the order of 92 Gt. This is of the same order of magnitude as the inferred increase in mass loss due to changes in ice dynamics over the past decade (Rignot and Kanagaratnam, 2006). The differences suggest that mass budget calculations (where the solid ice flux is subtracted from the SMB to estimate the overall net mass balance of the ice sheet) may have been seriously hindered by uncertainties in the SMB in the past. The most recent mass budget calculation, however, showed good agreement with results from independent satellite observations, giving confidence that progress is being made in constraining the components of mass loss from the ice sheet (van den Broeke et al., 2009). Nonetheless, resolving and explaining the cause of the large differences between reconstructions remains an important goal in improving the ability to model past, and predict future, changes in SMB.

To achieve this goal will require a concerted effort to collect more targeted in situ SMB data, especially from the percolation and ablation zone, with which to calibrate and validate the models. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow and subgrid-scale effects will also be valuable.

8.2.3. Ice discharge

- Accelerating ice discharge from outlet glaciers since 1995–2002 is widespread and has gradually moved further northward along the west coast of Greenland with the global warming.

- The ice discharge has increased from the pre-1990 value of 300 Gt/y to 400 Gt/y in 2005.

- The acceleration of fast-flowing marine terminating glaciers coincides with terminus retreat.

- The increase of glacier velocities is believed to be related to warming ocean water and the basal melt under the floating tongues can be several metres per day.

The Greenland Ice Sheet loses mass by melting as well as by discharge of ice into the ocean. The discharge comprises calving of icebergs as well as submarine melting. This section addresses the mechanisms of fast ice flow leading to transport of ice from the inland to the margin and the mechanisms of iceberg calving and submarine melting.

The margin of the Greenland Ice Sheet is characterized by spatially variable flow; areas of slow flow are separated by fast-flowing outlet glaciers and ice streams (Figure 8.15; Weidick, 1995; Rignot and Kanagaratnam, 2006). Parts of the ice are in direct contact with the surrounding ocean. In southern Greenland, ice-ocean contact is restricted to fast-flowing outlet glaciers that are topographically constrained by fjords, which often extend subglacially into the present-day ice sheet. The slower flowing ice usually terminates on dry land inland from the coast. However, in southwestern Greenland ice marginal lakes are quite common, and ice can be lost there by freshwater calving (Warren, 1991). The role of calving into ice marginal lakes has not been investigated extensively. It is likely to become increasingly important as the Greenland Ice Sheet retreats into the bedrock depression created by its own weight. In the north, along a significant section of the ice margin, both slowly and rapidly flowing ice is in direct contact with the ocean.

Around the Greenland Ice Sheet, the highest rates of ice discharge and the highest variability of discharge are observed at ocean-terminating outlet glaciers (Figure 8.15). The major mass loss mechanism for these glaciers is calving into the ocean (Reeh, 1968, 1994) and melting from ocean water (Rignot, 1996; Rignot et al., 1997). Rignot and Steffen (2008) showed that the floating ice shelf in front of Petermann Glacier, in northwestern Greenland, experiences massive bottom melting that removes 80% of its ice before calving into the Arctic Ocean. Commonly, estimates of ice discharge into the ocean implicitly include mechanical calving as well as melting by the ocean, since the ice contributes to sea level as soon as it crosses the grounding line and begins to float. Many of the outlet glaciers with high discharge rates flow through deep and narrow fjords. The largest and best studied of these is Jakobshavn Isbræ on the west coast of Greenland, for which the mechanisms of fast flow have been examined in detail (Iken et al., 1993; Funk et al., 1994; Lüthi et al., 2002, 2003). The findings are summarized here to illustrate the mechanism of fast outlet glacier flow.

Jakobshavn Isbræ is characterized by a deep subglacial trough that extends to 1500 m below sea level about 50 km inland from the current ice front (Clarke and Echelmeyer, 1996). This fjord (i.e., the trough) is relatively narrow (~ 5 km) but channels the ice from a drainage area that comprises about 6.5% of the Greenland Ice Sheet (Bindschadler, 1984; Echelmeyer et al., 1991). Ice flow in such a channel is inherently complicated. The various commonly-used ice flow models that are simplified by approximations based on shallow ice (i.e., where ice thickness is much less than horizontal extent) are not adequate to fully describe the flow in this type of channel. Such models include the Shallow Ice Approximation (SIA, Hutter, 1983), and the Shallow Shelf Approximation (SSA, MacAyeal and Thomas, 1982), which can also be applied to shallow ice streams with low basal drag (MacAyeal, 1989). Instead, a fully 3-D thermo-mechanical treatment of ice flow, without approximation of the stress terms in the momentum balance, is necessary (Hutter, 1983) to describe outlet glaciers whose depths are comparable to their widths. No such model has yet been applied to either the Greenland or Antarctic ice sheets or even to an entire drainage basin of one of these ice sheets. (The hierarchy of ice flow models is discussed in Section 8.3.2.)

The relatively steep and very thick ice of Jakobshavn Isbræ creates high driving stresses that are balanced by high basal drag, particularly at the flanks of the subglacial trough (Lüthi et al., 2003; Truffer and Echelmeyer, 2003). The high driving stress influences the stress field far into the surrounding ice sheet. Using a numerical model, Lüthi et al. (2003) found differences to sheet flow at lateral distances from the ice stream of seven times the ice sheet thickness. The high stresses lead to high deformation rates, because they are combined with soft ice at depth, due to a temperate layer and lower viscosity Wisconsian ice (Funk et al., 1994; Lüthi et al., 2002; Truffer and Echelmeyer, 2003).

The presence of this temperate ice adds an additional modeling challenge, because it introduces a free boundary: the cold-temperate surface (Hutter, 1982) and it requires the treatment of waterflow in the temperate layer (Aschwanden and Blatter, 2009). Also, the existence of a temperate basal layer implies water at the bed and a certain amount of basal motion. This is a boundary layer process that needs to be parameterized through a 'sliding law'. The most commonly used sliding laws (see also Section 8.3.2) relate

the basal motion to the basal shear stress and the subglacial water pressure; the latter is generally unknown but thought to be relatively high, that is, comparable to the ice overburden pressure.

This type of sliding law has well-known limitations, which are due to spatially and temporally variable parameters and to the limited knowledge of spatial and temporal variability of basal water pressure, which is an important variable (e.g., Iken and Truffer, 1997). The present inability to predict the nature and the amount of basal motion under large changes in geometry, water supply, and stresses is a major limitation to credibly predictive models of outlet glaciers.

Reeh (1994) used mass balance considerations to estimate a total ice discharge into the ocean of 316 Gt/y. This is higher than a subsequent estimate of 170 to 270 Gt/y, which is based on calving rate estimates from parameters such as ice thickness at the front and water depth (Biggs, 1999). Both estimates applied for some time before the 1990s. Reeh's (1994) estimate is very close to one by Rignot and Kanagaratnam (2006) for 1995, which was derived from measurements of ice surface velocity and thickness at flux gates, as well as an estimate of surface ablation downstream from these gates. Rignot and Kanagaratnam (2006) also estimated discharge for 2000 and 2005, which showed a progressive increase in discharge within just a few years (Figure 8.16). The various estimates are summarized in Table 8.3. It is clear that ice calving into the ocean is a significant ice loss mechanism for the Greenland Ice Sheet, and that it can be highly variable (Bamber et al., 2007). Ice calving is a major factor accounting for the recent negative mass balance of the Greenland Ice Sheet.

Table 8.3. Calving flux estimates for the Greenland Ice Sheet.

Calving flux estimates, Gt/y	pre-1990	1995	2000	2005
Reeh (1994) [a]	316			
Biggs (1999) [b]	170-270			
Rignot and Kanagaratnam (2006) [c]		321	354	421

[a] Estimate is based on mass balance considerations; [b] estimate based on empirical calving rate relationships derived in Greenland and elsewhere; [c] estimate derived from measurements of surface velocity and ice thickness at flux gates and a correction for surface ablation downstream of these gates. The measurements have been converted to Gt/y assuming a density of 900 kg/m^3.

The variability of ice discharge is confirmed by observations of individual outlet glaciers, which have shown dramatic changes in recent years. In the period 2002 to 2007, Jakobshavn Isbræ went through a period of thinning, retreat by loss of its floating tongue, and acceleration to twice its former speed (Thomas et al., 2003; Joughin et al., 2004, 2008c; Podlech and Weidick, 2004; Luckman and Murray, 2005). There is no evidence of similar increases in discharge in the 20th century at Jakobshavn Isbræ (Weidick and Bennike, 2007), but old moraines and glacier trim lines do indicate that the glacier has lost its floating tongue at least once since the Little Ice Age (15th to 19th centuries), while it was retreating to a position that remained stable for the next 50 years (Csatho et al., 2008). Large changes have also been observed at two big glaciers on Greenland's east coast: Kangerdlugssuaq Glacier and Helheim Glacier (Howat et al., 2005, 2007; Luckman et al., 2006; Stearns and Hamilton, 2007).

Rignot and Kanagaratnam (2006) showed that the acceleration of ice discharge is widespread. Between 1995 and 2000 many of the outlet glaciers south of 66° N accelerated, often doubling their discharge. This pattern spread to 70° N by 2005, and appears to have been related to regional warming. The detailed pattern is complicated, however. For example, Howat et al. (2007) reported that the Helheim and Kangerdlugssuaq glaciers decreased their discharge again to near former levels in 2006. This was partially due to a slow-down connected with a re-advance of the terminus, and partly due to thinning, so that at similar ice velocity less volume can be discharged. Other glaciers, such as Jakobshavn Isbræ, have maintained their high rates of ice discharge (Dietrich et al., 2007; Amundson et al., 2008; Joughin et al., 2008c).

The out-of-balance flux of ice into the ocean from many of the outlet glaciers in southern Greenland has led to large thinning of near-margin ice (Pritchard et al., 2009, Figure 8.17). This thinning is affecting large parts of the drainage basins (Stearns and Hamilton, 2007; Joughin et al., 2008c). For Jakobshavn Isbræ, the

thinning is closely related to increased ice flow that can be accounted for by an inland-propagating change in ice surface slope (Joughin et al., 2008c).

Coincident with ice-flow acceleration is a widespread retreat of outlet glacier terminus positions (e.g., Figure 8.18). Moon and Joughin (2008) recorded terminus positions of 203 outlet glaciers and concluded that retreat accelerated during 2000 to 2006 compared to 1992 to 2000. Again, the detailed picture is more complicated and Moon and Joughin suggested that there is a close relationship between air temperature and ice front advance / retreat, without suggesting that there is a cause and effect relationship between the two.

The observed acceleration of outlet glaciers coincides with an increase in teleseismically discovered glacial earthquakes (Ekström et al., 2003, 2006). The earthquakes occur seasonally and are strongly related to the occurrence of large calving events (Joughin et al., 2008b). Ekström et al. (2003) and Tsai and Ekström (2007) attributed the seismicity to large slip events on outlet glaciers. Recent ice displacement measurements do not confirm this hypothesis (Amundson et al., 2008; Nettles et al., 2008), and the former authors favor a model in which the glacial earthquakes are generated by the scraping of large overturning icebergs. Only a few seismic studies have been carried out on the ice and they point to a rich variety of seismic sources, many of which are presently not well understood (Rial et al., 2009).

Various hypotheses have been proposed for the causes of the speed-up of outlet glaciers. Thomas (2004) and Johnson et al. (2004) presented a model of Jakobshavn Isbræ in which the floating tongue provides resistance (i.e., backstress) to ice flow. The loss of the floating tongue then leads to ice flow acceleration. This model is also invoked by Joughin et al. (2008c) who observed a striking correlation between the speed of Jakobshavn Isbræ and the seasonally varying position of the calving front. Howat et al. (2005) used a force balance method to show how the loss of grounded ice near the terminus of Helheim Glacier leads to flow acceleration. The recently observed increases in ice discharge were always coincident with the loss of a floating tongue. But once the process is initiated, a pattern of fast flow, thinning and retreat can become established. This is a common observation at temperate tidewater glaciers, which only rarely develop floating tongues (Meier and Post, 1987). The idea of rapid tidewater glacier retreat is directly linked to that of a reverse bed slope, that is, the glacier bed deepens in the up-glacier direction. Nick et al. (2009) suggested that, in the case of Helheim Glacier, the phase of accelerated flow may be relatively short-lived, but it remains to be seen how universally applicable this model is.

While the exact form of a calving law remains unknown (e.g., Benn et al., 2007), it is clear from observations that a retreat into deeper water will lead to higher rates of calving, thinning, and further retreat. Vieli et al. (2001) and Schoof (2007) showed that no stable terminus position can exist on a reverse bedrock slope. This holds in advance as well as in retreat. The glacier acceleration that accompanies retreat can then be due to the effect of thinning, as suggested by Meier and Post (1987) and used by Pfeffer (2007).

Thinning influences glacier velocity in several ways. First, spatially-uniform thinning leads to a reduction in driving stress and thus in horizontal velocity. Second, thinner ice leads to a reduction in effective stress (ice overburden minus basal water pressure). This is because the basal water pressure has to exceed a level that is given by sea level for subglacial drainage to occur. As a tidewater glacier thins, basal water pressure cannot decrease due to this set downstream level, causing an overall drop in effective pressure. The reduction of effective pressure, as the lower part of the glacier approaches floatation, is expected to lead to higher rates of basal motion in any reasonable sliding law. Finally, enhanced thinning near the terminus leads to increases in surface slope and hence driving stress and ice velocity.

Pfeffer (2007) derived a stability index for tidewater glaciers by considering the first and second of the three effects, which have opposite effects on ice flow. Pfeffer showed that an ice thickness above a certain threshold:

$$h_{th} = C \cdot \rho_w / \rho_i \, h_w$$

is stable against thinning in that thinning leads to a reduction in ice flow, whereas an ice thickness below that threshold reacts to thinning by ice acceleration. Here, h_w is the depth of the water and ρ_w and ρ_i are the densities of ocean water and glacier ice, respectively. C is a constant that depends on the choice of the ice flow law and the sliding law. Pfeffer (2007) suggested that C = 4/3, given flow parameters in the literature.

The analyses by Vieli et al. (2001), Pfeffer (2007) and Schoof (2007) all apply to flowline models and do not consider the embedding of an outlet glacier in its surrounding ice sheet. This could be an important factor contributing to the stability of outlet glaciers. As an outlet glacier retreats and thins, the surface slope increases in a lateral direction (i.e., toward the ice sheet), and more convergent flow should be expected, which could stabilize the glacier and delay further thinning and retreat (Howat et al., 2007).

Measurements from outside the fast-flowing ice in West Greenland have also indicated a seasonality in ice flow (Lüthi et al., 2002; Zwally et al., 2002b; Joughin et al., 2008b; van de Wal et al., 2008). This seasonality is related to the seasonality of surface melt and was attributed to meltwater reaching the base of the ice (Zwally et al., 2002b). Das et al. (2008) observed the draining of a supraglacial lake to the bed and a corresponding local speed-up of the ice sheet. They concluded that the integrated effect of multiple lake drainages could explain some of the observed acceleration of the ice sheet. However, for several reasons, it is unlikely that increased melt directly leads to the observed speed-up of outlet glaciers.

First, the ice-stream acceleration was first observed near the termini of these glaciers and then propagated inland (Joughin et al., 2004; Howat et al., 2007). Second, at least in the case of the main channel of Jakobshavn Isbræ, there is no observed distinct seasonal reaction to the large variability of meltwater supply. Rather, the seasonality of ice flow is closely related to the position of the glacier terminus, which varies seasonally, but with a distinctly different phase than the meltwater production (Joughin et al., 2008c). Third, van de Wal et al. (2008) did find a large seasonal velocity signal that correlates well with meltwater production, but does not translate into a correlation between annual velocities and melt. In fact, the mean annual velocity of the ice sheet has a negative correlation with melt. This has also been reported for mountain glaciers (e.g., Truffer et al., 2005) and is presumably due to the theoretically expected increase in effectiveness of water drainage as discharge increases.

Reeh et al. (2001) and Joughin et al. (2008c) observed a correlation between sea-ice cover and the stability of glacier fronts. While it is perhaps difficult to imagine a sea-ice cover that provides substantial resistance to glacier flow, it is possible that icebergs held together by sea ice inhibit calving, which then leads to glacier advance because ice is not lost at the terminus (Joughin et al., 2008c). This can then have an influence on ice velocities by the increased resistance due to lateral shear, and perhaps by decreased effective pressures near the grounding line due to ice thickening. In practice, it is often difficult to tell whether the influence of sea ice is direct, or whether the sea ice reacts to the same forcing as the outlet glaciers, for example, warmer ocean water.

That all large observed changes in ice-stream flow initially occurred near the ocean and then propagated inland, points to the importance of understanding ice-ocean interactions. Holland et al. (2008) reported that the acceleration of Jakobshavn Isbræ is temporally coincident with a pulse of deep warm ocean water that was observed along the west coast of Greenland. Typical fjord circulation patterns that bring in warm saline water at depth and return colder and fresher water at the surface could bring this pulse to the glacier front and contribute to the very high melting rates that were reported on Jakobshavn Isbræ before the disintegration of its floating tongue (Thomas et al., 2003). Straneo et al. (2010) and Rignot et al. (2010) also pointed to the importance of warm water incursions for regulating glacial flow. Howat et al. (2008) noted a correlation between sea surface temperature and outlet glacier acceleration and retreat in southeastern Greenland. Murray et al. (2010) observed simultaneous acceleration and deceleration of many southeastern glaciers and proposed that increased ice and freshwater discharge led to a strengthening of a fresh coastal current that kept warm ocean water from entering glacial fjords, thus providing a negative and self-regulating feedback between ice discharge and melting by ocean water. Ocean conditions can also favor the formation of a strong melange (mix of sea ice and icebergs) that can inhibit calving and lead to a strong seasonality in terminus position (Amundson et al., 2009). The length of the winter season determines the amount of time a grounded terminus is exposed to calving. This provides a link between seasonality and longer term glacier behavior.

In a different setting, Motyka et al. (2003) showed that melting of submerged ice at a tidewater glacier front could account for about half of the total ice flux into the ocean. The important factors for melting were the temperature of the deep fjord water and the strength of the fjord circulation. Motyka et al. (2003) hypothesized that the buoyant freshwater emerging at the glacier's base plays an important role in the fjord

water circulation. This provides an interesting physical link between atmospheric temperatures, surface melt, and fjord circulation.

The fast flow of Jakobshavn Isbræ and other outlet glaciers can be explained as a consequence of subglacial topography. This is not, however, necessarily the case for all outlet glaciers. One that stands out in particular is the northeastern Greenland ice stream (Fahnestock et al., 1993; Figure 8.15), which extends far into the ice sheet (Joughin et al., 2000). The fast flow of this ice stream might be controlled by geothermal heat. Fahnestock et al. (2001) used the spacing of deep radar layers to derive the geothermal heat flux under this ice stream and discovered unusually high melt rates.

Some of Greenland's glaciers also show surge-type behavior: long periods of quiescent flow interrupted by short periods of very rapid flow (Meier and Post, 1969). Glacier surges are a non-equilibrium phenomenon that is believed to be due to a positive feedback between basal water storage and basal motion (Kamb et al., 1985; Harrison and Post, 2003) that can last several months. Slower surges at polythermal glaciers might also be thermally controlled (Clarke et al., 1984). In Greenland, surging has been reported for the northeastern ice stream Storstrømmen (Reeh et al., 1994) and several glaciers in central East Greenland (Murray et al., 2002; Jiskoot et al., 2003). A short surge-like flow acceleration was also seen at Ryder Glacier (Joughin et al., 1996). It is important to remember the temporary nature of such changes in ice discharge. The mechanisms (and perhaps duration) are quite different from outlet glacier changes.

8.2.4. Total mass balance

- The Greenland Ice Sheet is losing mass at an accelerating rate.
- The present [5-year average or 2005/06] loss of mass is 205 Gt/y corresponding to a 0.5 mm/y global average rise in sea level.
- Satellite and airborne observation in general reveal a slight increase in elevation in the regions above 2000 m and a strong decrease in elevation in the regions below 2000 m with very dominating regional patterns.

8.2.4.1. Total net balance of the Greenland Ice Sheet

The total mass balance of an ice sheet is the sum of the surface mass balance and ice discharge. It is the net result of the mass changes due to the addition and loss processes discussed in the previous sections. Owing to the seasonal nature of these processes, the mass of an ice sheet varies throughout the year, between years, and over many years as longer-term trends in climate become established. There are strong spatial gradients in mass balance that must be taken into account to determine variations in the total mass of the ice sheet.

For the Greenland Ice Sheet, which is subject to substantial surface melt and the resulting runoff of meltwater, mass exchange through the year has a large seasonal component at lower elevations caused by winter accumulation giving way to summer surface melt. These changes around the ice sheet margins are accompanied in outlet glaciers by changes in ice flow and iceberg production that may be linked in part to summer warming. At high latitudes and high elevations in the interior of the Greenland Ice Sheet, the relatively low input of snow and relatively small loss of snow to sublimation and wind transport lead to smaller amplitude annual variations in mass.

The time it takes for significant changes in accumulation or rates of ice loss around the margins of the Greenland Ice Sheet to impact on the ice flow in the bulk of the ice sheet interior determines the response of the ice sheet to forcings on ice-age time scales. Changes in the flow of the Greenland Ice Sheet over millennia are driven by slowly changing temperature profiles and ice characteristics. The rate of these changes is controlled by thermal diffusion, slow flow of ice downward and outward as new snow is added to the ice sheet surface, and the slow evolution of the shape of the ice sheet as changing accumulation and melt patterns add and remove mass.

Prior to recent measurements of changes in the rates of outlet glacier flow and melt in Greenland, the ice sheet appeared to be roughly in balance. The mass turnover in the Greenland Ice Sheet was estimated at 500

Gt/y by Benson (1962) with about half the mass added through snowfall each year lost by surface melt and subsequent runoff, and the other half returned to the ocean through outlet glacier discharge leading to iceberg calving. But, as discussed in Sections 8.2.2 and 8.2.3, recent observations show quite large and rapid changes in surface melting and ice discharge. At the same time, different analyses of the total mass balance show the Greenland Ice Sheet overall to have lost mass since the early 1990s.

8.2.4.2. Techniques employed to observe or estimate total mass balance

Current techniques for estimating or measuring total mass balance include: (i) the mass budget approach, comparing total net snow accumulation with losses by ice discharge and meltwater runoff; (ii) repeated altimetry, to estimate volume changes; and (iii) tracking temporal changes in gravity, to infer mass changes.

The first two provide estimates of mass balance for regions included within survey boundaries, whereas the third provides an estimate for large regions with less distinct boundaries. All three techniques can be applied to the Greenland Ice Sheet as a whole, within limitations specific to each technique.

8.2.4.2.1. Mass budget approach

Changes in the Greenland Ice Sheet mass can be inferred from the difference between estimates of accumulation (i.e., snowfall) and mass loss through ablation (i.e., melt and subsequent runoff and sublimation) or ice discharge and calving. Each of these quantities involves a number of variables. Generally, snow accumulation is estimated from annual layering in ice cores, sometimes with interpolation between core sites using satellite microwave measurements or shallow radar sounding (e.g., Rotschky et al., 2004), or from atmospheric modeling (e.g., Box et al., 2004). Ice discharge is the product of ice flow velocity and ice thickness, with velocities measured in situ or remotely, preferably near the grounding line where velocity is almost depth independent. Ice thickness is measured by airborne radar, seismically, or from measured surface elevations assuming hydrostatic equilibrium for floating ice near grounding lines. Generally, meltwater runoff (large near the Greenland coast but small or zero elsewhere) is from model estimates calibrated against surface observations where available (e.g., Box et al., 2004; Hanna et al., 2005). Rignot and Kanagaratnam (2006) applied the mass balance approach around Greenland.

Mass budget calculations involve the often small difference of two large numbers, and small errors in either mass loss or gain terms can result in large errors in estimated total mass balance. These errors are difficult to assess for the Greenland Ice Sheet because of high temporal and spatial variability (see Section 8.2.2). Although broad interferometric SAR (InSAR) coverage and progressively improving estimates of grounding line ice thickness have substantially improved ice discharge estimates, incomplete data coverage and uncertainty regarding velocity-depth relationships and meltwater runoff limit accuracy of total discharge estimates. The mass budget uncertainty for the Greenland Ice Sheet is a result of all these factors.

Moreover, accumulation estimates are based on data from the past few decades, and there are indications that snowfall in Greenland may be increasing with time (e.g., Hanna et al., 2005). Similarly, it is becoming clear that glacier velocities can change substantially over quite short time periods (see Section 8.2.3). Both of these would add to estimated mass budget errors.

Thomas et al. (2000) gave a regional estimate for the mass budget for central Greenland above 2000 m elevation, using discharge velocities from repeat GPS measurements made at stake markers 30 to 40 km apart along a traverse at 2000 to 3000 m elevations, ice thickness from radar sounding, and snow accumulation rates from shallow ice cores. Results show that between the mid-1970s and mid-1990s, central Greenland was in overall mass balance, but with regions of thickening, particularly in the southwest, and of thinning, particularly in the southeast.

A more complete picture, including near-coastal regions and individual drainage basins, comes from balancing net SMB (Section 8.2.2) against ice discharge rates derived from SAR-derived ice velocities and ice thicknesses from radar or surface elevations near grounding lines, where the ice is floating (Rignot and Kanagaratnam, 2006). SMB for this work was derived from SMB anomalies calculated from Hanna et al. (2005) scaled and applied to modulate a surface accumulation pattern from shallow ice cores combined with ECMWF (re)analyses (Section 8.2.2).

Results show significant mass loss since the mid-1990s from most drainage basins, with very large losses after 2000 from Jakobshavn Isbræ, Kangerdlugssuaq Glacier and Helheim Glacier and from nearly all glaciers along the east coast south of Helheim (Joughin et al., 2004; Rignot and Kanagaratnam, 2006; Luckman et al., 2006; Howat et al., 2007, 2008). Velocity time series for these glaciers show up to a 100% increase after 2000, consistent with their negative mass balance.

8.2.4.2.2. Repeat altimetry

Rates of surface elevation change (δS/δt) reveal changes in ice sheet mass after correction for changes in depth / density profiles and bedrock elevation, or for hydrostatic equilibrium if the ice is floating. Satellite radar altimetry (SRALT) data have been widely used (e.g., Davis et al., 2000; Johannessen et al., 2005; Zwally et al., 2005), together with laser altimetry from airplanes (Krabill et al., 2000), and from NASA's ICESat (Zwally et al., 2002a; Thomas et al., 2006). Modeled corrections for isostatic changes in bedrock elevation (e.g., Peltier, 2004) are small (i.e., a few mm/y) but with errors comparable to the correction. Those for near-surface snow density changes (Arthern and Wingham, 1998; Li and Zwally, 2004; Helsen et al., 2008) are larger (i.e., a few cm/y) and also uncertain.

SRALT data provide the longest time series, with the first measurements in 1978. Resulting estimates of δS/δt can be misleading, however, because derived elevations refer to the radar reflection horizon. The depth of this horizon below the actual ice sheet surface is affected by near-surface characteristics, such as snow wetness. Consequently, changes in these characteristics can affect SRALT-derived estimates of δS/δt (Thomas et al., 2008), particularly in Greenland where the area affected by summer melting has increased substantially since the early 1990s. In addition, because of the broad radar beam, SRALT data are unreliable over the sloping and undulating surfaces near the coast, where rates of elevation change are greatest (e.g., Brenner et al., 2007).

Airborne Topographic Mapper (ATM) and satellite (ICESat) laser altimeters provide data that are easier to validate and interpret because footprints are small (i.e., about 1 m for airborne laser, and 60 m for ICESat) and there is negligible laser penetration into the ice. However, clouds limit data acquisition, and accuracy is affected by atmospheric conditions and particularly by laser pointing errors. Moreover, existing laser data are comparatively sparse, being limited by practical limitations on the number of ATM surveys and the ICESat orbit separation of up to 40 km in southern Greenland.

ATM elevation estimates are accurate to ~10 cm along survey tracks (Krabill et al., 2002) with similar errors estimated for ICESat surveys, but increasing over steeper slopes (Brenner et al., 2007). Because there are large gaps both in ICESat and airborne coverage, δS/δt values are generally supplemented by degree-day estimates of anomalous melting (Krabill et al., 2000, 2004). This increases overall errors and probably underestimates total losses, because it does not take into full account the dynamic thinning of outlet glaciers.

In summary, δS/δt errors cannot be quantified precisely for SRALT data, because of the broad radar beam and time-variable penetration. For laser data the difficulty in quantification comes because of sparse coverage. If the SRALT limitations are real, they are difficult, if not impossible, to resolve. Laser limitations result primarily from poor coverage, and can be resolved by increasing coverage to include all regions of rapid change around the Greenland coast. All altimetry mass balance estimates include additional uncertainties due to changes in the density profile of near-surface snow and (small) rates of basal uplift. If elevation changes are caused by recent changes in snowfall, the appropriate density (ρ) may be as low as 300 kg/m^3; for long-term changes, it may be as high as 900 kg/m^3. This is of most concern for small rates of surface elevation change, where the safest assumption is that density $\rho = 600 \pm 300$ kg/m^3, implying a ±50% mass balance uncertainty. Slobbe et al. (2009) used this conservatively large range for the entire ice sheet for an ICESat-only analysis, finding results consistent with other laser analyses.

Large and rapid changes, commonly found near the coast, are almost certainly caused by changes in melt rates or glacier dynamics, and the appropriate density of the lost or added volume approaches that of pure ice (ρ ~900 kg/m^3). Moreover, densification rates are sensitive to snow accumulation rates, temperature, and wetness, with warm conditions favoring more rapid densification (Arthern and Wingham, 1998; Li and Zwally, 2004; Helsen et al., 2008). Thus, the recent Greenland warming probably caused ice sheet surface

lowering simply from this effect alone. Finally, the rate of basal uplift is inferred from models and has uncertain errors.

An early estimate of surface elevation change in southern Greenland using SRALT showed either little change or that much of the variability could be attributed to decadal changes in accumulation rates as measured in ice cores (Davis et al., 2000; McConnell et al., 2000). Other estimates of the Greenland Ice Sheet volume change from SRALT altimetry (Johannessen et al., 2005; Zwally et al., 2005) show thickening at higher elevations, particularly in the south, comparatively modest thinning nearer the coast, and a small positive mass balance overall.

However, as previously discussed, time-variable radar penetration and 'blurring' of near-coastal results by the wide radar beam make these results suspect (Thomas et al., 2008). Estimates derived by comparing ATM surveys 1993/1994 and 1998/1999 (Krabill et al., 2000) show substantial thinning near the coast and an overall mass loss of about 55 Gt/y between surveys. Comparison of elevations from these surveys with later values from ICESat at locations where flight tracks crossed ICESat orbits show that mass losses have increased to about 80 Gt/y between 1998/1999 and 2004 (Thomas et al., 2006). These laser-based estimates probably represent lower bounds because the sparse coverage 'misses' outlet glaciers that are thinning dynamically. Annual thinning rates of tens of metres on outlet glaciers in southeastern Greenland are shown clearly from repeat stereo satellite coverage (Howat et al., 2007, 2008; Stearns and Hamilton, 2007).

8.2.4.2.3. Temporal variations in Earth's gravity field

Of all the techniques that are used to monitor the changing mass of the Greenland Ice Sheet, only gravity measurements are directly sensitive to changes in that mass; all other techniques convert changes in other quantities, such as melt, ice flux or surface elevation change, into changes in mass. Since 2002, the GRACE (Gravity Recovery and Climate Experiment) mission, joint between NASA and Deutsche Forschungsanstalt für Luft und Raumfahrt (DLR), has measured Earth's gravity field and its temporal variability. GRACE comprises two satellites, following each other in the same orbit 200 km apart, that carefully monitor the distance between themselves (Tapley et al., 2004a,b).

As the first satellite approaches a more massive object on the surface, it accelerates away from the trailing satellite, which then accelerates as it approaches the same object, closing the increased distance between the two orbiters. A detailed record of the changes in this inter-satellite distance over time provides a direct indication of changes in mass at the Earth's surface. Other accelerations are monitored with on-board accelerometers. GRACE has produced time series of these small changes in gravitational acceleration that start in 2002 and still continue (Tapley et al., 2004a,b). The spatial scale of the changes in gravity that GRACE can detect is determined by the height of the orbit, the spacing of the two satellites, and the accuracy with which the range rate between the two satellites can be determined.

After removing the effects of other loading, high latitude data, like those for Greenland, contain information on temporal changes in the mass distribution of the ice sheet and the underlying bedrock. Because of its high altitude, GRACE makes coarse-resolution measurements of the gravity field and its changes with time and the resulting mass balance estimates are also at coarse resolution – hundreds of kilometres. Despite this limitation, this technique has the advantage of covering the entire Greenland Ice Sheet, which is extremely difficult using other techniques. GRACE estimates include mass changes on the small ice caps and isolated glaciers that surround the Greenland Ice Sheet, regions not included in mass balance estimates from other techniques.

There are two different analysis schemes for GRACE data that have been applied to determining changes in the mass of the Greenland Ice Sheet. The first uses products derived by the GRACE mission that represent the global gravity field, expressed as a spherical harmonic expansion to a high order, differencing fields from each 30-day period to detect changes in ice sheet mass over time (Velicogna and Wahr, 2005, 2006; Chen et al., 2006). This technique requires convolving the harmonic expansion of the global gravity field with a filter representing either part or all of the Greenland Ice Sheet to isolate the changes in gravity that are due to changing ice mass. The gravity fields produced by the GRACE team have been updated since the time of publication of these papers; results may change with this revised input data. Ramillien et al. (2006) used a similar technique, but began with a different input dataset, which is a version of the Earth's geoid (i.e.,

another representation of the gravity field) derived for 10-day periods by a group at Centre National d'Etudes Spatiales (CNES). Slobbe et al. (2009) used a common analysis scheme on gravity fields produced by four different centers, and reported a significant range of resulting net balance rates.

Corrections applied in these analyses include compensation for changes in global patterns of water storage on land, ocean mass, and atmospheric mass. In addition, a correction must be made for longer-term changes in crustal mass distribution due to glacial isostatic adjustment (GIA) (crustal uplift that is a continued response to past deglaciation, or post-glacial rebound). Each of these analyses uses different input data sets and modeling strategies to make the corrections. Chen et al. (2006) used an optimized filtering technique to fit mass change time series on a 1° × 1° grid with high resolution aided by use of knowledge about where changes were occurring (i.e., in outlets near the ice margin identified by Rignot and Kanagaratnam, 2006). Ramillien et al. (2006) discussed the impacts of errors from each of these corrections. Barletta et al. (2008) conducted an extensive analysis of the impact of different factors on the GIA corrections to GRACE estimates in Greenland and Antarctica. They did make a somewhat simplified estimate of the rate of mass loss in Greenland from GRACE data to illustrate these impacts. This result is not included in Figure 8.19 or Table 8.4, but is similar to the others in the table.

Table 8.4. Rate of total mass change in the Greenland Ice Sheet.

Source	Technique	Time period	Rate of total mass change, GT/y	Range (or stated error), ± Gt/y	Notes
	Radar Altimetry				Satellite radar altimetry is affected by surface slope near ice margins and changes in penetration
Zwally et al., 2005	SRALT (ERS 1-2) + limited ATM	1992 – 2002 (10.5 y)	11	± 3	
Johannessen et al., 2005	SRALT (ERS 1-2) low elevations sparsely sampled	1992 – 2003 (11 y)	30	± 3	Densities of 0.9, 0.5, 0.4, 0.3, 0.3 Gt/km^3 assumed to convert area-weighted height change in elevation intervals (<1.5, 1.5–2, 2–2.5, 2.5–3, >3 km) into mass
	Laser Altimetry				
Krabill et al., 2004	ATM (airborne laser altimetry)	1993/4 – 1998/9	-55	± 3	
		1997 – 2003	-73	± 11	
Thomas et al., 2006	ATM	1993/4 – 1998/9	-4 to -50		High end of range likely due to limited coverage at low elevations
	ATM/ICESat	1998/9 – 2004	-57 to -105		High end of range likely due to limited coverage at low elevations
Slobbe et al., 2009	ICESat (GLAS)	2003/2 – 2007/4	-139 ± 68		Used large density range (± 300 kg/m^3) for error bounds
	Mass Budget				
Rignot and Kanagaratnam, 2006	InSAR (ice motion) 1996 + balance anomaly estimate (Hanna et al., 2005)	1996	-83	± 28	
	InSAR (ice motion) 2000 balance anomaly estimate (Hanna et al., 2005)	2000	-127	± 28	
	InSAR (ice motion) 2005 + balance anomaly estimate (Hanna et al., 2005) (extrapolated)	2005	-205	± 38	
	Satellite Gravity (GRACE)				

Luthcke et al., 2006	GRACE MASCON – (change in gravitational acceleration on each orbital pass fit to forward model of mass history of local basins)	July 2003 – July 2005	-101	± 16	Gain of 54 GT/y above 2000 m and loss of 155 GT/y below 2000 m
Chen et al., 2006	GRACE – averaging filter applied to monthly spherical harmonic gravity field product	April 2002 – Nov 2005	-219	± 21	
Velicogna and Wahr, 2006	GRACE – averaging filter applied to monthly spherical harmonic gravity field product	April 2002 – April 2006	-227	± 33	
Ramillien et al., 2006	GRACE – averaging filter applied to 10-day gravity solutions	July 2002 – March 2005	-118	± 14	
Wouters et al., 2008	GRACE EOF decomposition of monthly spherical harmonics iteratively fit with forward model of mass change of basins (based initially on R&K and iterated)	February 2003 – January 2008	-179	± 25	
		July 2003 – July 2005	-121	± 27	Same period as Luthcke et al. (2006) Similar elevation dependence seen
		January 2006 – January 2008	-204	± 25	
Slobbe et al., 2009	GRACE using products from CNES (Centre National d'Etudes Spatiales), CSR (Center for Space Research, University of Texas-Austin), DEOS (Delft Institute of Earth Observation and Space Systems, Delft University of Technology, Delft), and GFZ (GeoForschungsZentrum, Helmholtz-Zentrum Potsdam)	2002/7 – 2007/6	-128 to -218		Range in net balance is range of results using gravity products from four different centers, rather than estimated error

Wu et al. (2010) have presented a different analysis also based on the glacial isostatic adjustment and the redistribution of ice and water masses. It is suggested that the GIA correction is large and negative (-0.72 ±0.006 mm/y) resulting in a mass loss of 104 Gt annually during the period 2003 to 2008 from the Greenland Ice Sheet; this mass loss is smaller than those published by other authors.

The second type of GRACE analysis scheme uses the range rate measurements between the two satellites directly, modeling the local changes in mass required to produce the observed changes in these rates during different overpasses of Greenland (Luthcke et al., 2006). This work also makes corrections for GIA using published paleo ice sheet reconstructions, but not the smaller effects from atmospheric and oceanic mass variations.

Error sources can include measurement uncertainty, leakage of gravity change signal both from regions surrounding the Greenland Ice Sheet, and for some analyses, globally, and causes of gravity changes other than ice sheet changes. Of these, the most serious are the gravity changes associated with vertical bedrock motion (i.e., GIA). These are inferred from models of deglaciation history and its impact on vertical crustal motion, primarily post-glacial crustal uplift; the errors are also modeled and remain highly uncertain. Results from all gravity-based analyses show mass loss over their periods of observation, with rates ranging from -100 Gt/y to more than -200 Gt/y. The wide spread in results requires further investigation, but several studies using different analysis techniques show continued and accelerated change over the GRACE time period.

8.2.4.3. Synthesis and state-of-the-art

The results for each of the techniques are shown in Figure 8.19, plotted as boxes that encompass the length of the measurement period along the time axis, and the published range of the mass change rate over that interval. Data used for Figure 8.19 are summarized in Table 8.4.

With the exception of measurements from SRALT, estimates of the Greenland Ice Sheet total mass balance discussed here are negative, and are becoming more negative with time, with the caveat that many of these records are relatively short. The trend and timing are largely consistent with increased ice discharge from outlet glaciers around the edge of the Greenland Ice Sheet. Where these measurements do not agree within published errors, the sources of the discrepancies may be attributable to shortcomings in spatial sampling, complex sets of corrections, and input data that continue to evolve as satellite missions mature; limitations in measurement techniques are better characterized, and surface conditions on the Greenland Ice Sheet respond to a changing climate.

The complexity of measuring or estimating total balance has led to the development and exploitation of a number of measurement strategies. With respect to the Greenland Ice Sheet, these strategies have been developed and are maturing into a system that will one day be capable of tracking well constrained and internally consistent mass changes on a sub-annual basis. The extension of present measurement time series, combined with follow-on instruments and new airborne and satellite campaigns (e.g., Cryosat II), will help solve problems associated with limited coverage, slope and surface change induced error, and disparities in analysis of common data sets.

This represents a rapid evolution in capability that is presently matched by the rapid evolution of the Greenland Ice Sheet mass itself. One lesson that may be taken from the present state of knowledge is that multiple approaches and observations of mass input and mass loss processes have been required to make the progress that has been made to date.

8.2.5. Summary and outlook

Compared with processes that take place at the base of the ice sheet, those that influence SMB are relatively well understood and considerably easier to observe. Despite this, published reconstructions of the individual components that make up SMB differ significantly. The differences can be as large as the inferred increase in mass loss due to changes in ice dynamics over the past decade. The uncertainties in SMB arise, largely, as a result of the paucity of relevant, spatially extensive in situ observations. To reduce uncertainties to a level useful for mass budget calculations and reliable SMB

predictions will require a concerted effort to collect more targeted in situ data, especially from the percolation and ablation zone. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow and sub-grid scale effects will also be needed. Although there are differences between reconstructions, they all indicate a negative trend over the past ~15 years. This is largely due to an increase in runoff since about 1995, resulting in a marked reduction in SMB since then. Also of note is that 2007 had the lowest SMB of any year over the 50-year record for which reliable estimates exist.

Much of the ice loss from the Greenland Ice Sheet occurs by ice discharge into the surrounding ocean. This has been estimated at 50% of the mass loss on average over the past ~50 years, but recently large variability in the flow of many large outlet glaciers has been documented. The temporal variability of Greenland's ice mass balance has a large contribution from variability of flow in its outlet glaciers. Observations during the past decade have shown that ice discharge can increase by a factor of two within a few years and, in some cases at least, that this can also be reversed. These changes are coincident with observations of warming in the ocean and atmosphere, as well as the disappearance of near-coastal sea ice. While the general mechanisms of fast flow are reasonably well understood, the necessary tools for predicting future behavior have not been developed. This is primarily due to a lack of understanding of the ice-ocean coupling and the role of surface water in determining the rates of ice flow.

The total mass balance of the ice sheet (i.e., the sum of SMB and ice discharge) was exceptionally difficult to determine prior to the development of the present suite of observational and analysis techniques. While it is clear from the spread of published results that estimation remains a challenge for these techniques, it is important to note that nearly all the approaches show very similar trends, indicating clearly that the Greenland Ice Sheet is losing significant mass, and has been doing so at an accelerating rate over the past ten years. This conclusion is supported by all approaches sensitive to the ice margins: (i) the change in ice discharge from outlet glaciers combined with increasingly negative SMB; (ii) surface lowering in outlet glaciers measured directly by laser altimeters; and (iii) the reduction in the mass of the ice sheet measured by the GRACE satellite gravity mission. This reduction of mass is consistent with what would be expected to happen in a warming Arctic and the diverse, independent suite of approaches used to document this change in total mass balance provides confidence in the result.

The challenge of more accurately determining the present and future rate of mass change of the Greenland Ice Sheet, by reducing the uncertainties inherent in the different approaches, will require improvements in measurements, more consistent and complete observational time series, better analysis schemes, and an improved understanding of the physical processes involved in recent rapid changes.

This chapter has attempted to show the state-of-the-art with respect to observations and understanding of present-day, and recent past, behavior of the Greenland Ice Sheet. It is clear that, although the state-of-the-art has greatly advanced over the past decade or so, major gaps and challenges remain. During this period, unpredicted and striking variability in ice dynamics has been detected from satellite observations as well as a marked decrease in SMB coincident with a decreasing trend in total mass balance. Serious limitations in the ability to model and monitor these processes, however, are also apparent from the disparity between published results.

Progress will need to come from many sources, including:

- Improving acquisition of more comprehensive and complete time series suitable for ice velocity determination, including seasonal variability. This will be predominantly from satellite and airborne remote sensing.
- Acquisition of improved measurements of surface elevation change by satellite and aircraft.
- Better measurement of ice thickness in rapidly flowing glaciers for improved ice discharge estimation.
- Improvements in analysis strategies and measurement capabilities for all processes which determine SMB.
- More, and better, in situ measurements of refreezing and ablation.

- Better understanding of links between outlet glaciers, surface melt, and marine conditions.
- Perhaps, most importantly of all, the need to ensure that the time series of observations is secure into the future, and extended back in time with the aid of historic and paleo-proxy data.

8.3. Predictions and sensitivity

- The Greenland Ice Sheet is likely to lose mass at an accelerating rate in the 21st century in response to ongoing warming.

- Increases in snow accumulation in central Greenland may partially offset the increased melting and runoff from the ice sheet periphery, but are unlikely to compensate for the effects of higher temperatures, which increase the area of the ablation zone as well as the duration and intensity of the summer melt season.

- Between 5 and 10 cm of global sea level rise can be expected by 2100 as a result of declining SMB. Adding estimates of increasing ice discharge means a total ice mass loss of 10 to 19 cm can be expected by 2100.

- Greenland Ice Sheet retreat is projected to accelerate beyond 2100 under most IPCC emissions scenarios, as the regional climate continues to warm and the dynamical feedbacks associated with surface lowering become more important over time scales of several centuries.

- A regional warming of 3 to 4.5 °C (equivalent to a global warming of 2 to 3 °C) would move the Greenland Ice Sheet into a state of negative SMB, which would cause large-scale ice sheet collapse if maintained for several centuries to millennia.

The Greenland Ice Sheet response to climate change is too complex to be predicted through simple theory or analytical equations. SMB depends on the interaction of weather systems with the ice sheet orographic influences and surface energy balance. Spatial and temporal patterns of ice sheet evolution therefore feed back onto the meteorological fields over the ice sheet. The regional- and synoptic-scale weather patterns that affect the ice sheet are also evolving as part of global-scale climate change and accompanying shifts in sea ice, ocean conditions, the hydrological cycle, and atmospheric dynamics. These systems are all interlinked and the interactions can be described only by numerical weather or climate models (Box 8.1).

Box 8.1. Numerical models of the oceans, atmosphere and ice sheets

Numerical models of the oceans, atmosphere, and ice sheets take similar approaches to simulating the large-scale evolution of each system. All are based on the principles of mass, momentum, and energy conservation, with these conservation equations governing mass and energy transfer within or between these systems. The models dissemble the oceans, atmosphere, and ice sheets into 3-D grid-cell networks and solve the governing equations to predict the system state in each grid cell. Neighboring cells interact to simulate the fluxes of water, air, ice, energy, and climatically active variables such as salt and water vapor.

The whole system is driven by boundary conditions such as global topography and bathymetry, the Earth's rate of rotation, and temporal and spatial patterns of solar input. The resulting fluid motions and energy exchanges provide a representation of the climate system. Integrating the system forward in time, system interactions can give internal weather/climate variability or can simulate climate system response to a change in boundary conditions, for example, increases in solar activity or greenhouse gas concentrations.

The resolution of the simulation is determined by the size of the 3-D grid cells used to discretize the oceans, atmosphere, and ice sheets. Although computational limitations are continually receding, simulating climate dynamics still taxes the world's most advanced computational facilities. This places limits on the available simulation capabilities. For century-scale climate change scenarios,

typical grid resolutions are 1° latitude-longitude for the oceans, 2° to 3° for the atmosphere, and 0.2° (~20 km) for ice sheets.

Similarly, the evolution of the ice sheet depends on the climate fields that dictate SMB (see Section 8.2.2), on the oceanic influences that affect the rates of iceberg calving and basal melting at the ice-ocean interface, and on the dynamic ice sheet processes that govern the flux of ice to the ice sheet margin. Numerical models have been developed to describe the ice sheet. However, some processes are poorly understood and unresolved in the models, particularly those associated with basal sliding of ice over the bedrock. Furthermore, SMB and ocean-ice interactions need to be prescribed to simulate ice sheet response to climate change. Although efforts to couple ice sheet, atmosphere, and ocean models to describe the co-evolution of these systems are still in early stages, this is the best available route for physically-based forecasts of how the Greenland Ice Sheet will change in the coming decades and centuries.

8.3.1. Aspects of modeling climate change for Greenland

The IPCC AR4 models incorporate many improvements compared to their predecessors, but there are still shortcomings (see Stroeve et al., 2007; as well as Chapter 2). While some studies suggest anthropogenic forcing may favor a positive Northern Annular Mode (NAM), there is evidence that climate models underestimate NAM-like variability (e.g., Stenchikov et al., 2006). Maslanik et al. (2007) analyzed three types of atmospheric circulation pattern that appear most significant in terms of Arctic Basin winds and ice transport. The 'light ice' phases of these patterns include decreased mean sea level pressure (SLP) in the North Atlantic (an 'NAO-like' pattern resembling the positive phase of the NAO), a low pressure cell within the Arctic Basin (a 'central Arctic' pattern), and a dipole pattern of high pressure over the Canadian Arctic paired with low pressure over the Siberian Arctic. Winds and ice transport patterns that favor reduced ice cover in the western and central Arctic have continued since the late 1980s, but the Arctic Oscillation (AO) index is not a reliable indicator of these patterns.

Atmospheric modeling over complex terrain is a great challenge. The major reason for this is the difficulty represented by the topographical relief. If not properly resolved, the atmospheric circulation will not, for an otherwise perfect model, be able to depict the real local circulation patterns. This, in turn, will lead to systematic deficiencies in the models' ability to capture regional / local climatological features.

The most obvious example is that of topographically lifted precipitation that causes enhanced precipitation on the upstream slopes of a mountain, while producing less precipitation on the lee side downstream. Only when the appropriate mountain ridges are captured by the resolution, will the precipitation patterns be accurately simulated. The importance of this effect at larger scales is not well understood.

High resolution is offered by regional climate models (RCMs) (see Chapter 2). Results from RCMs are sensitive to the choice of integration domain, the horizontal and vertical resolution and parameterizations, and all contribute to the uncertainties of RCM simulations. There is no exact choice of domain any more than there is for resolution or parameterization, which means that domain must be treated as yet another element contributing to the uncertainty of RCM simulations. RCM simulations are also sensitive to lateral boundary conditions and to the surface conditions prescribed. The climate scenario projections come from interpolated atmosphere-ocean general circulation model (AOGCM) simulations and contribute to the overall uncertainties of RCM simulations and projections. The RCM results can vary depending on the AOGCMs used, in terms of large-scale atmospheric circulation patterns and their teleconnections. RCM results are often based on only one specific AOGCM and are for just one selected scenario. A multi-model ensemble approach could increase the confidence of future projections of extreme climate. Accurate simulation of surface mass accumulation over the ice sheets requires a spatial resolution which is currently not available from AOGCM simulations. Therefore high-resolution RCMs (i.e., PMM5 and MAR) have been run for shorter time slices driven by AOGCM boundary conditions and combined with an ice sheet mass balance model as described by Gregory and Huybrechts (2006) and Fettweis et al. (2008), see also Chapter 2.2.

One of the remaining challenges is that sequential modeling (i.e., climate modeling followed by ice sheet modeling) lacks any possible feedback between the climate and ice sheet. Thus, a poor climate model performance over the ice sheet will result in a possibly erroneous climate change signal over the ice sheet as well. This in turn will possibly result in an erroneous estimate of the change in SMB, even if the ice sheet model is excellent. A consistent development and coupling of models of atmospheric processes with ice sheet models is therefore in demand (see also Chapter 2).

According to the recent IPCC AR4 GCM studies, the Arctic is projected to be generally warmer and wetter (Kattsov et al., 2007). Using an RCM, Rinke and Dethloff (2008) projected future changes over the Arctic by the end of the 21st century according to the IPCC A1B emissions scenario. The results showed distinct regional patterns. Over the Arctic Ocean these are mainly associated with sea-ice retreat and over land are mainly associated with surface characteristics, such as orography, and the agreement with the large-scale temperature and precipitation projections within the driving GCM is striking. Regional detail is evident along the North Atlantic storm track and, during summer, close to orographic obstacles. This is attributed to the higher resolution of the RCM.

Modeling sea ice in a realistic manner is still a great challenge, particularly with respect to the minimum ice extent at the end of summer. Dorn et al. (2007, 2008, 2009) investigated modified descriptions of ice growth, snow and ice albedo, and snow cover on ice used in the coupled regional atmosphere-ocean-ice model HIRHAM-NAOSIM. A series of sensitivity experiments were performed in order to assess the need for more sophisticated parameterizations of these processes in coupled regional and global models. These showed that the simulation of Arctic summer sea ice is very sensitive to the parameterization of snow and ice albedo as well as to the treatment of ice growth. The parameterization of the snow cover fraction on ice plays an important role in the onset of summer ice melt. This has a crucial impact on summer ice decay when more sophisticated schemes for ice growth and ice albedo are used. The work by Dorn and co-workers illustrates that when using a harmonized combination of more sophisticated parameterizations, simulation of the summer minimum in ice extent can be considerably improved due to the more realistic representation of interactions between the atmosphere and sea ice influencing Greenland and the Arctic as a whole.

Current understanding of model limitations as briefly described here, suggest that as a necessity for achieving improved simulations of the climate over Greenland, model development efforts should be directed towards the following:

- Use of an accurate ice sheet topographic dataset
- Accurate albedo parameterization
- More realistic surface layer model, including thermal conductivity for sno /firn
- Accurate planetary boundary layer momentum exchange.

These recommendations would improve the simulation of regional effects of Greenland's climate due to finer-resolved orography and land-sea contrasts, better-resolved non-linear interactions between the large-scale and mesoscales, better simulation of hydrodynamic instabilities and synoptic cyclones, and better description of hydrological and precipitation processes. Better representation of the northward development of the storm track, and invasion of the Arctic Basin by cyclonic systems from the lee of Greenland, which varies greatly with the NAO, would have a major influence on the Arctic sea-ice movement and melting in a coupled regional climate system. At higher resolution, the air-sea interaction changes, the air-sea heat flux is enhanced and localized in the lee of Greenland, and the wind stress increases.

8.3.1.2. Model limitations and challenges

In order to capture even the large-scale snow accumulation rates, it is essential for the atmospheric models to include a good representation of the mountains around the Greenland Ice Sheet. The detailed deposition of moisture in the mountains as well as over the ablation zone of the ice sheet will influence the amount of moisture transported further into the interior of the ice sheet. It is also of great importance to use an accurate elevation model for the ice sheet itself. Box and Rinke (2003) looked at how the ice sheet topography used in the standard version of the HIRHAM regional climate model compared to a more recent digital elevation model (e.g., Bamber et al., 2001), and found substantial

differences. They considered these to be responsible for several degrees of systematic temperature bias, particularly in the ablation zone.

Figure 8.20 illustrates how the topography of Greenland is depicted with increasing resolution (based on Bamber et al., 2001). At the coarsest grid resolution (~150 km), which is typical for most advanced state-of-the-art GCMs, the resolution does not allow for discrimination between the outermost part of the ice sheet and the coastal mountains. This basically implies that the atmosphere 'sees' all of Greenland as one big lump of ice. Increasing the grid resolution to 25 km improves the situation, but the gradients over the ice margins are not resolved even at this scale. Typically, the incremental rise in topography within the ablation zone is 300 to 500 m for a 25 km grid, which is potentially in conflict with the formulation of the vertical structure of the atmospheric model and clearly introduces systematic surface temperature errors that may exceed 3 °C. At a grid resolution of 12 km, this feature is almost resolved. But it is not until a resolution of about 5 km is applied that the surface temperature error no longer introduces serious inconsistencies, in cases where the modeled temperature field is used without any adjustment.

At present, it is still speculative as to whether a grid resolution of 5 km will also improve the precipitation climatology. Figure 8.21, however, compares the precipitation patterns from three simulations with the RCM HIRHAM with increasing resolution (Stendel et al., 2008; Philippe Lucas-Picher, Danish Meteorological Institute, pers. comm. 2008). The three HIRHAM realizations have a grid distance of ~80 km, 25 km and 5 km, respectively. The large-scale patterns are similar, but whether the precipitation is falling in the coastal margins, in the ablation zone or even in the accumulation zone does not appear to be resolved unless using a fine grid resolution.

In order to provide realistic driving conditions for an ice sheet model, a necessary requirement for the atmospheric model must, therefore, be that it resolves the features that are of most relevance in controlling the mass balance of the ice sheet. If this cannot be provided directly, additional adjustments to the data are necessary. This, in turn, raises questions about internal consistency of the energy and water cycle balances.

Representing the detailed regional conditions of Greenland in a climate model is also affected by the ability of the model to simulate realistically the energy and moisture cycles in the surrounding seas. Clearly, part of the moisture supply for the precipitation systems influencing the ice sheet originates from the regional seas surrounding Greenland. Because the oceanic circulation around Greenland is complex and poorly resolved by coarse-resolution coupled climate models, it is not surprising that detailed sea-ice conditions are often poorly represented.

At best, only large-scale sea-ice conditions can be depicted. Stendel et al. (2008) gave an example of a climate change downscaling experiment, where the driving GCM does have a relatively good representation of present-day sea-ice conditions, both with respect to seasonal variability and to interannual variability. However, details are still not matching observed conditions to a level providing confidence that the changes are indeed a result of a realistic timing of changes in circulation and moisture support.

When ice sheet models are forced with atmospheric model output from GCMs, it is always an issue as to how realistic the simulated sea-ice temperatures and sea-surface temperatures actually are.

The IPCC concluded in its Fourth Assessment Report (IPCC, 2007) that mean temperature within the Arctic is likely to increase at a greater rate than mean global temperature, confirming results from previous IPCC assessments and the Arctic Climate Impact Assessment (ACIA, 2005). For the IPCC A1B emissions scenario, the annual mean Arctic temperature increase by the end of the 21st century is projected to be about twice that of the global mean increase (5 to 7 °C *vs* 2.5 to 3.5 °C) (IPCC, 2007; Christensen et al., 2007). Figure 8.22 shows an extract from an evaluation of 21 model simulations of global change under the A1B emissions scenario (Christensen et al., 2007), highlighting three models, the NCAR PCM, GFDL CM2.0, and MPI ECHAM5 as well as the 21-model mean. It is clear that the three individual models qualitatively show the same climate change response, but that the magnitude

of the change differs by several degrees with the PCM model showing the lowest and the ECHAM5 model the highest degree of warming, while the GFDL CM2.0 model is close to the ensemble mean.

It is interesting to note, however, that the projected climate signals are to some degree caused by quite different mechanisms. Figure 8.23 shows an extract of an analysis of the performance of 14 model simulations for the period 1958 to 2000 (Walsh et al., 2008). Again, the ensemble mean behavior is shown with the same three individual models. A common feature for most of the models, reflected by the ensemble mean, is a clear cold bias in the Barents Sea due to a tendency to simulate too much sea ice, with the MPI model a clear exception. The implication is that only the MPI model simulates sea-ice coverage in this region reasonably well. At the same time, the greatest warming by the end of the study period is simulated exactly over this region in the ensemble mean as well as by the individual models.

In the NCAR and GFDL models, this partly reflects the bias in present-day sea-ice conditions, while this apparently cannot be the case in the MPI model because the present-day sea ice appears to be captured with some realism. It should also be noted that, in general, the greatest warming occurs in the area with too much ice (strong cold bias) under current conditions, and in the NCAR model in particular, even though here winter data are compared with the annual mean.

This section has shown that, to varying extents, results at the regional scale are clearly subject to systematic errors in present-day simulations. Using an ensemble of models masks this deficiency. The role of the ocean as a driver of the climate in the Arctic is obvious, and there should be concern about interpreting a climate change signal, when most of the signal is apparently due to systematic error in simulating present-day conditions.

As the models discussed above have indicated, maps of warming must be carefully analyzed and, without further analysis, data cannot be used in a region with non-linear feedbacks such as the presence and absence of sea ice. Thus, it is also likely that the simulated warming and precipitation change signals are strongly influenced by these systematic errors, not only in the vicinity of the imperfections themselves, but also further away due to non-linear interactions in the atmospheric and oceanic circulations.

8.3.2. Modeling Greenland Ice Sheet dynamics

8.3.2.1. Objectives of numerical ice sheet modeling

The future evolution of the Greenland Ice Sheet depends on the extent of climate change over the ice sheet, in particular the specific impacts on SMB as well as past climate forcing. As discussed in Sections 8.2.3 and 8.2.4, overall ice sheet mass balance depends strongly on the rate of ice discharge. In particular, dynamical mass loss is governed by the flux of ice to the ablation area. This includes terrestrial margins, where ice is removed through melting and runoff, and marine margins, where ice loss occurs through iceberg calving and melting of ice that is in contact with the ocean. High ice velocities lead to greater transport of ice to the ice sheet margins and higher rates of mass loss through each of these processes.

Ice sheet dynamics are difficult to predict or extrapolate from present-day observations because ice sheet velocities do not remain constant on decadal and century timescales. There are unanswered questions as to how ice flux will respond to changes in ice sheet geometry, oceanic boundary conditions, and surface meltwater reaching the bed. All of these changes are expected in the coming decades and centuries.

Numerical ice sheet models have been designed to describe first order ice sheet physics and to provide prognostic solutions for ice sheet evolution in response to such changes. Numerical models are required because the flow and deformation of ice at a particular point depends on a complex array of physical parameters, including ice sheet thickness, surface and bed slopes, ice temperature (which influences the deformation of the ice), ice fabric, valley walls, regional ice flow, and basal friction which is a function of bed geology, basal ice temperature, and subglacial hydrological conditions. By

simulating these processes and features of the ice sheet system, physically-based estimates of ice flow and deformation and thus ice sheet response to climate change are possible.

8.3.2.2. The physical basis of ice sheet models

Detailed descriptions of ice sheet physics are provided by van der Veen (1999) and Paterson (1994). Box 8.2 provides a simplified overview of the theoretical basis of ice sheet models. The ice sheet is discretized into a 3-D array of grid cells and the equations that govern ice thickness, velocity, and temperature evolution are solved in each grid cell. Atmospheric and oceanic forcing of the ice sheet are introduced as boundary conditions.

Box 8.2. Theoretical basis of ice sheet models

Three-dimensional models applied to the Greenland Ice Sheet to date all make use of the shallow-ice approximation, in which the gravitational driving stress is locally balanced by drag at the glacier base (Nye, 1957; Hutter, 1983). The gravitational driving stress is $\tau d\ (z) = \rho\ g(hS - z)\ \nabla hS$, where ρ is the ice density, g is gravitational acceleration, hS is the glacier surface elevation, and ∇hS is the surface slope. At the glacier bed, $\tau d\ (hb) = \rho\ gH\ \nabla hS$, for ice thickness H. Glen's flow law relates ice deformation rates to the stress field in the ice. In the shallow-ice approximation, the ice velocity (u) associated with vertical shear deformation (d), i.e., ud, follows $ud \propto \nabla hS\ n\ H\ n+1$, where n = 3 is the exponent in Glen's flow law.

Ice deformation rates are sensitive to ice temperature, with the effective viscosity of ice varying by a factor of about 1000 over the range of temperatures found in the Greenland Ice Sheet (Marshall, 2005). Ice sheet models therefore simulate the 3-D ice thermodynamics – advection and diffusion of energy and strain heating due to ice deformation – to model the temperature distribution in the ice sheet.

In addition to internal deformation, ice can flow via basal motion where the bed is at the pressure-melting point, through some combination of subglacial sediment deformation and decoupled sliding over the bed. Basal flow rates, ub, often exceed the ice motion associated with deformation by one or several orders of magnitude. The vertically-averaged velocity, $\bar{u}$, is calculated from the sum of these two contributions: $\bar{u} = \bar{u}d + ub$. Large-scale basal flow generally requires a layer of pressurized subglacial meltwater at the bed. Models make some allowance for basal flow, usually through a local sliding 'law' relating basal flow rates to gravitational shear stress at the bed. There is little knowledge of subglacial hydrological conditions, and the physical processes that determine basal water pressure are not yet modeled or parameterized in ice sheet models. For this reason, sliding rates and the spatial-temporal variability in basal flow are highly uncertain in current models.

Given an estimate of the vertically-averaged ice velocity, the equation for conservation of mass describes the rate of change of ice thickness at each point on the ice sheet:

$$\frac{\partial H}{\partial t} = -\nabla \cdot (\bar{u} H) + b$$

The first term on the right describes the divergence of the ice flux and b is the local mass balance rate (accumulation minus ablation; basal melting can be included in this term but is, generally, negligible compared to surface accumulation and ablation). This is usually considered over one year, that is, expressed as metres per year of ice-equivalent gain or loss of mass. Given measurements or climate model predictions of b, this equation can be integrated forward to simulate the evolution of ice thickness at all locations on the ice sheet. This is the basis of ice sheet modeling.

The Greenland Ice Sheet is still adjusting to the last cycle of glaciation and deglaciation, particularly with respect to ice temperature at depth, where most of the ice deformation occurs. Simulations of future evolution of the ice sheet therefore require a 'spin up' simulation that takes Greenland through one or more glacial cycles (Huybrechts et al., 1991; Letréguilly et al., 1991). The climate forcing for

these simulations is taken from Greenland ice core records. Climate forecasts are then imposed for studies of future ice sheet changes, generally as perturbations from present-day climatology.

When considering longer time scales, adjustment of the Earth's crust to ice loading and unloading must be included in the model as this represents a potentially important feedback through the coupling between SMB and surface elevation (e.g., Oerlemans, 1980). Different models have been adopted for estimating isostatic adjustment, with for example, Marshall et al. (2000) assuming a local, damped return to isostatic equilibrium, while Le Meur and Huybrechts (1998) incorporated a self-gravitating spherical visco-elastic Earth model.

For whole-ice sheet simulations over one or more glacial cycles the horizontal grid spacing is typically 20 km or more, with some 30 layers in the vertical (e.g., Huybrechts, 2002). Grid sizes of 5 to 10 km are tractable, but each time the grid is reduced by a factor of two the computational demand increases by an order of magnitude. There are also theoretical and pragmatic limits to resolution. The shallow-ice-approximation that is used in most whole-ice sheet models is based on an assumption that horizontal grid dimensions are greater than the ice thickness. Further, available bedrock maps for the ice sheet, derived from airborne radar surveys, are currently available only at 5-km resolution. Issues associated with model resolution are discussed below.

Parameterization of SMB is a key parameterization in the models. Most ice sheet modeling efforts to date prescribe the SMB from monthly or annual surface air temperatures and precipitation rates. These can be based on observed (present-day) meteorological patterns or meteorological fields downscaled from a climate model. In general, the SMB fields simulated by climate models are not accurate enough for coupled ice sheet-climate modeling, in part because GCMs are too coarse to adequately resolve the ice sheet ablation areas (see Section 8.3.1). This makes it difficult to estimate ablation through a direct surface energy balance, so degree-day methods, driven solely by temperature fields, are generally adopted for melt and runoff modeling, following Reeh (1991) and Huybrechts et al. (1991).

Interest in decadal-scale ice sheet changes and the development of improved regional-scale meteorological models (e.g., Box et al., 2006) should soon permit direct estimates of SMB from meteorological models for future predictions, rather than parameterizations based on temperature and precipitation fields. This is desirable because such models are more physically-based and can capture the spatial variability in climate fields as ice sheet geometry and surface properties change; parameterizations tuned to present-day conditions become less reliable as both the climate and the ice sheet depart from their present-day state. This is discussed in more depth in Sections 8.2.2, 8.3.1, and 8.3.3.

8.3.2.3. Modeling of the Greenland Ice Sheet

To date, Greenland Ice Sheet simulations are all based on the shallow-ice approximation and the physics described in Box 8.2 (Huybrechts et al., 1991, 2004; Letréguilly et al., 1991; Greve, 1997, 2000; Ritz et al., 1997; Huybrechts and de Wolde, 1999; Marshall and Cuffey, 2000; Tarasov and Peltier, 2002; Gregory and Huybrechts, 2006). Basal flow is typically assumed to be active wherever the basal ice is at the pressure-melting point and is prescribed as a function of the gravitational driving stress.

Figure 8.24 gives an example of modeled and observed topography for the Greenland Ice Sheet on a 5-km grid. Present-day ice sheet area and volume can be simulated to within a few percent, but there are often systematic regional biases in modeled ice thickness (Figure 8.24c). The Northeast Greenland Ice Stream (see Figures 8.15 and 8.16) is present in the model (Figure 8.25b), but the modeled flow rates in the ice stream are too slow, contributing to the ice thickness anomaly in this region. The ice margin position is poorly simulated in northern Greenland, leading to thickness anomalies of several hundred metres. The model is anomalously thin in southern-central Greenland, for reasons that are unclear. It may be related to a bias in the climate forcing or modeled rates of basal flow that are excessive, driving a thinner, lower-sloping ice dome.

The velocity field over the whole ice sheet indicates high flow rates on the ice sheet flanks and areas of concentrated ice flux where they are expected (Figure 8.16). These regions correspond to major topographic channels and drainage outlets, which are resolved in this simulation. However, the model fails to capture the high rates of discharge in major outlet glaciers such as Jakobshavn Isbræ and Helheim Glacier; simulated velocities in some of these systems are an order of magnitude too low. This is most likely to be related to inadequate resolution of the fjord geometry, discussed further below.

Basal flow in this model is parameterized as a function of gravitational driving stress and is enabled wherever the ice sheet bed is at the pressure-melting point. Figure 8.25 plots modeled present-day basal ice temperatures and ice sheet velocities, including contributions from basal flow in warm-based sectors of the ice sheet. Few data are available to validate or test these predictions from the model.

Deep borehole data confirm that central and eastern Greenland are generally cold-based, due to the relatively thin ice that drapes the underlying mountains and the advection of cold ice to the bed along the ice divide. On the ice sheet flanks, deformational heating warms the ice to produce large areas that are at the melting point. In the major topographic channels, this can produce a thick temperate ice layer which has a low effective viscosity, introducing a positive feedback that accelerates deformation and ice flow in these channels.

These models have limitations, elaborated below, but they also capture many large-scale features of the ice sheet dynamics in Greenland. Bearing the model limitations in mind Figure 8.26 illustrates the modeled response time of the Greenland Ice Sheet to simple warming scenarios. A regional warming of 2 to 8 °C is introduced as a linearly increasing temperature perturbation over the period 2000 to 2100, relative to late 20th century baseline temperature fields over the Greenland Ice Sheet as compiled by Ohmura (1987). The temperature anomaly is held constant from 2100 to 3000. Figure 8.27 plots ice sheet configurations at 3000 for the 4 and 8 °C scenarios.

Most of the ice volume loss in these scenarios comes from southwestern Greenland, with the ice dome in southern Greenland collapsing under sufficient warming. This result has been well established in the modeling by Huybrechts and others, as illustrated in Figure 8.28 from Alley et al. (2005). These results correspond to three different scenarios of future atmospheric carbon dioxide (CO_2), with values in excess of 750 ppm inducing summer warming of more than 5 °C in Greenland, sufficient to excite collapse of Greenland's southern dome on a timescale of several centuries.

These results reflect the plausible impact of atmospheric warming on increased melt and drawdown of the Greenland Ice Sheet. There is an accelerating ice loss in late stages of the retreat, when the interior of the Greenland Ice Sheet falls below the equilibrium line altitude. Due to simplified fast-flow and ice-marginal physics in the model, as well as one-way climate forcings (i.e., missing climatic feedbacks of ice sheet retreat), the time scale of modeled Greenland Ice Sheet retreat in Figures 8.26, 8.27 and 8.28 is uncertain. Many of the missing feedbacks and processes give a systematic bias towards underprediction of ice sheet sensitivity to climate warming; it is expected that Greenland Ice Sheet retreat will proceed more quickly than forecast for a given warming scenario. However, how quickly this could transpire is not well-constrained, because no models with the requisite physics are available to assess this. This is picked up again in Section 8.3.3. The next section addresses several of the outstanding challenges for the current ice sheet models.

8.3.2.4. Modeling challenges

The modeling studies cited above have a general skill in simulating Greenland Ice Sheet dynamics, but some features and processes are difficult to capture. These include ice sheet-ocean interactions, ice stream dynamics, basal flow processes (sliding and sediment deformation at the glacier bed), and coupling of ice marginal dynamics and inland ice dynamics. These features are critical to questions of ice sheet sensitivity and response time to climate change.

The Fourth Assessment Report of the IPCC summarizes this well:

Dynamical processes related to iceflow not included in current models but suggested by recent observations could increase the vulnerability of the ice sheets to warming, increasing future sea level rise. Understanding of these processes is limited and there is no consensus on their magnitude. (IPCC, 2007:17).

An expanded discussion of shortcomings of existing ice sheet models was provided by van der Veen and ISMASS (2007). Shortcomings with existing Greenland Ice Sheet models fall into three categories: those related to physical processes; those associated with spatial resolution; and uncertainties in specification of boundary conditions, including ice-ocean and ice-atmosphere interactions. These issues are elaborated in the following sections, ranked according to a subjective judgment of their relative importance to modeling Greenland Ice Sheet ice dynamics. These processes and considerations have different levels of importance in different parts of the ice sheet, and in some cases the process is not understood well enough to assess how important it will prove to Greenland Ice Sheet response to climate change.

8.3.2.4.1. Uncertainties associated with physical processes

H1_Basal flow and hydrology

The extent and physical controls of glacier motion due to basal flow remain poorly known and difficult to quantify (Clarke, 2005). This is true across a wide spectrum of ice masses, including mountain glaciers, outlet glaciers in Greenland and other polar icefields, and Antarctic ice streams. Basal flow is generally the main mechanism for fast flow (except for enhanced creep rates in deep fjords), so must be well quantified to provide a realistic model of ice sheet dynamics.

Elaborate theories have been constructed to describe the sliding of ice over variously shaped obstacles and glacier beds with different roughness characteristics, but it is unclear whether the models are relevant to large-scale basal flow. The models describe salient physical processes in the subglacial environment, including regelation and stress enhanced softening of glacier ice, which allow it to flow around basal obstacles, and the free slip of basal ice over air- and water-filled cavities in the lee of bedrock obstacles. While these processes are likely to contribute to integrated basal slip over large areas of the glacier bed, at least for hard beds (i.e., bedrock rather than deformable till), the governing physics operate at scales of centimetres to metres. Basal motion at scales of tens of metres to tens of kilometres, as observed in Antarctic ice streams and ice marginal regions of Greenland (Joughin et al., 2008a), is more relevant to overall ice flux.

Furthermore, motion associated with regelation and stress-softening around bedrock obstacles increases non-linearly with shear stress (Paterson, 1994:157), whereas basal flow in many settings (e.g., Antarctic ice streams) is associated with low values of gravitational driving stress: thin, low-sloping ice with subglacial water pressures near flotation, implying low values of basal shear stress over large regions (Bindschadler et al., 2000). Summer speed-ups in valley and outlet glaciers and the ice marginal region of the Greenland Ice Sheet provide another illustration of this. These speed-ups are associated with negligible change in surface geometry (i.e., gravitational driving stress), but are instead driven by reductions in basal friction due to the influx of surface meltwater to the bed (e.g., Iken and Bindschadler, 1986; Copland et al., 2003; Das et al., 2008; Joughin et al., 2008a).

The mechanics of large-scale basal motion therefore appear to be governed by the extent of the glacier bed that experiences ice-bed decoupling due to high subglacial water pressures. High water pressure promotes basal flow by effectively floating the glacier above the bed, reducing basal traction. This applies both to basal sliding and subglacial sediment deformation, although the rheological properties and supply of basal sediments also need to be considered for the latter.

The need to understand hydrological influences on basal flow and to incorporate these processes into models is pressing, considering recent evidence that surface meltwater can drain to the bed through hundreds of metres of cold ice in marginal areas of the Greenland Ice Sheet (Zwally et al., 2002b; Das et al., 2008) and smaller-scale polar icefields (Boon and Sharp, 2003). Similar to the long-standing observations from mountain glaciers, this influx of meltwater is capable of triggering transient speed-

ups, increasing the discharge of ice (Zwally et al., 2002b; Joughin et al., 2008a). This presents a direct mechanism by which climate change can influence ice dynamics. Coupled with trends of increasing spatial extent and intensity of the melt season in Greenland (Tedesco, 2007), accelerated summertime flow is expected in future decades in the marginal zones of the Greenland Ice Sheet.

The magnitude of the speed-up and the extent to which ice sheet drawdown will propagate inland are less clear. Joughin et al. (2008a) presented observations over a 300-km wide region in southwestern Greenland which indicate that summer speed-ups of the major outlet glaciers, including Jakobshavn Isbræ, contribute only a few percent to the total annual discharge. Speed-ups on the slow-moving ice sheet flank (i.e., away from the major outlet glaciers) are relatively greater, increasing annual discharge by as much as 25%, but this region does not account for much of the total dynamic discharge of the Greenland Ice Sheet. Therefore, this mechanism represents a pathway for ice sheet response to climate forcing, but one which may be relatively stable (i.e., non-catastrophic).

These influences on ice dynamics are generally absent in the large-scale models that predict Greenland Ice Sheet response to climate change. Physically-based models of ice sheet hydrology are needed to simulate these processes. In a flowline model of the western flank of the Greenland Ice Sheet, Parizek and Alley (2004) simulated the supraglacial hydrology and englacial drainage to the bed, assuming that surface drainage induces speed-ups as seen by Zwally et al. (2002b). Extrapolating their results to the whole ice sheet, their most conservative scenarios gave 12% to 38% more volume loss from the Greenland Ice Sheet over the next several centuries due to surface-meltwater induced lubrication, with the effect increasing with warming. Parizek and Alley (2004) also suggested that the effect could be much higher, with the upper range of their modeling giving a tripling of the ice volume loss in Greenland, relative to models that exclude surface meltwater effects.

These models assume that basal flow is proportional to water supply at the bed, but subglacial water storage and basal water pressures are what really matter for ice-bed coupling and basal flow. The physics that governs the relationship between subglacial water pressure, bed geology, bed roughness, and basal traction still needs to be elucidated and parameterized in a way that is physically valid and testable in models.

A subglacial hydrological model is also required to provide a realistic estimate of spatial and temporal variations in basal water pressure. These have begun to be developed and coupled with models of glacier dynamics (Arnold and Sharp, 2002; Flowers and Clarke, 2002; Johnson and Fastook, 2002; Flowers et al., 2005; Marshall et al., 2005), but subglacial hydrological conditions are notoriously difficult to observe, so models are difficult to test and constrain at large scales. To date, most models of subglacial hydrology treat the subglacial water system as a distributed water sheet, with water fluxes driven by hydraulic potential gradients. Conduit physics and the nuances of regime shifts between distributed and channelized drainage systems need to be better understood and included in the models in order to capture spatio-temporal patterns of basal water pressure and their influence on glacier dynamics. This has yet to be attempted in Greenland.

H1_High-order stresses (horizontal stress coupling)

The shallow-ice approximation estimates ice motion as a function of the local gravitational driving stress, with the explicit assumption that non-local effects are negligible. This is known to be a poor approximation in certain parts of the ice sheet, such as the ice divide, ice streams, ice shelves, and ice sheet margins. In these settings, higher-order stress terms such as horizontal shear stress and longitudinal stress gradients, sometimes known as non-local stress coupling effects, become important to ice deformation. For instance, horizontal shear stresses at ice stream margins and next to valley walls play an important role in holding back ice flow, while longitudinal strains (stretching) represent the primary mode of ice deformation in ice shelves and floating ice tongues.

Longitudinal stress coupling also provides a mechanism for the transmission of ice-marginal forcings to the interior of the ice sheet. Current thinning of Jakobshavn Isbræ is accompanied by rapid retreat of the calving terminus (Sohn et al., 1998; Joughin et al., 2004). Joughin et al. (2004) proposed that the velocity increase resulted from development of large rifts in the floating tongue, followed soon

thereafter by complete collapse of this tongue. Associated reductions in buttressing would allow increased discharge from the grounded portion of the glacier (Thomas, 2004). On the other hand, Pfeffer (2007) proposed that thinning could have decreased the height above buoyancy, thereby reducing the effective basal water pressure. This, in turn, would lead to greater sliding speeds, especially if the ice thickness approached floatation. To adequately incorporate these processes into numerical models and to test the validity of these competing hypotheses, it is necessary to incorporate stress transmission across the grounding line.

Models of glacier dynamics with higher-order stresses have become more common in studies of valley glaciers as well as regional ice sheet applications (Albrecht et al., 2000; Schneeberger et al., 2001; Pattyn, 2002, 2003; Saito et al., 2003). These advances are relatively new, and have yet to be applied to whole-ice sheet simulations in Greenland. Higher-order models such as that of Payne et al. (2004) and Price et al. (2008) demonstrate the importance of longitudinal stress coupling for accurate simulation of ice sheet response to perturbations at the ice sheet margin, that is, inland propagation of changes due to ice-marginal thinning or acceleration.

H1_Iceberg calving

Ice sheet grounding lines, floating ice tongues, and iceberg calving processes are poorly represented in current glaciological models. Existing models generally determine the position of the calving terminus (i.e., marine ice front) by either prescribing an ice limit based on a maximum water depth (e.g., the continental shelf break), or by freely simulating floating ice tongue advance based on a simple parameterization of iceberg calving, typically prescribing calving rates to be proportional to water depth. Zweck and Huybrechts (2005) examined several different treatments of calving in large-scale models; all are simple parameterizations that do not describe the actual process(es) and should be considered to be statistical-empirical models that aim to capture the bulk loss of ice through this mechanism.

The physical controls of iceberg calving are not well understood and may not be deterministic. Iceberg calving requires fractures to initiate and propagate at the ice front. The process has been well-observed, but the environmental controls of calving rates are difficult to isolate; ocean and air temperature, surface meltwater, sea-ice conditions, and tidal flexure at the ice front may all have an influence on calving, but there is no simple relationship. Sohn et al. (1998) observed an increase in calving rate following the onset of spring melting and found a strong correlation between surface melting on the Greenland Ice Sheet, the melting of fjord ice, and the rapid increase in calving rate during summer. Their observations are insufficient, however, to resolve conclusively whether the presence of abundant surface meltwater enhances the ability of surface crevasses to penetrate the ice thickness, or whether the break-up of confining fjord ice reduces the restraining force acting on the calving front.

Recent evidence indicates the likely role of regional ocean warming in destabilizing the floating ice tongue of Jakobshavn Isbræ (Holland et al., 2008). This is also consistent with tidewater-style retreat of Helheim Glacier on the southeast coast of Greenland (Nick et al., 2009). Ocean-ice sheet coupling is lacking in the current models, but clearly needs to be embraced for predictive models of ice sheet response to climate warming.

8.3.2.4.2. Uncertainties associated with model resolution

Whole-ice sheet models for Greenland now operate at scales of 5 to 20 km, but scale problems still arise because of the small spatial scale of most of the outlet glaciers where drainage to the sea is concentrated. For example, Jakobshavn Isbræ, which drains mass from 7% of the total area of the Greenland Ice Sheet, is less than 10 km wide over its entire ~100 km length (Echelmeyer et al., 1991). This glacier overlies a narrow (6 to 10 km) and deep (~1500 m below the surrounding basal landscape) subglacial trough (Clarke and Echelmeyer, 1996; Gogineni et al., 2006; Lohoefener et al., 2006) that cannot be resolved with a horizontal grid spacing of 10 km or more. This limits the capacity of models to simulate the enhanced creep and high rates of flow that are observed in Jakobshavn Isbræ and other major outlet glaciers that drain the Greenland Ice Sheet.

At the same time, detailed and accurate bed topography upstream of the grounding line or calving front is needed to assess the possible irreversibility of terminus retreat for Greenland's major outlet glaciers. Howat et al. (2007) suggested that the rapid retreat of the Helheim Glacier terminus occurred as the calving front retreated into deeper water and continued until reaching a reversed bed slope, similar to the behavior of tidewater glaciers (Nick et al., 2009).

Even if spatial resolution is improved sufficiently to capture small-scale basal topography, narrow channels no more than a few kilometres wide and up to 1500 m deep challenge the assumptions in the shallow-ice approximation. As noted by Clarke and Echelmeyer (1996), these bedrock troughs are similar to valley walls in that the near-vertical faces produce lateral drag that partially opposes the driving stress – horizontal shear stresses become important in taking up the gravitational driving stress. Also, an overlying temperate layer of fast-moving ice is embedded within more slowly moving ice, which may provide additional resistance, similar to the situation on West Antarctic ice streams. As a start, an effective shape factor could be introduced (Clarke and Echelmeyer, 1996) to account for the additional flow resistance, but no rigorous tests have been conducted to evaluate the applicability of this approach.

Additional uncertainty is introduced by the well-known (if rarely discussed) sensitivity of modeled ice margin dynamics to grid resolution (Abe-Ouchi and Blatter, 1993). With a fixed grid and a free boundary, the standard treatment in continental ice sheet models, the advance and retreat of the ice sheet margin proceed by discrete 'jumps' to adjacent grid cells. This process is sensitive to grid resolution because the flux of ice into a neighboring cell is a function of the local surface slope. This sensitivity combines with mass balance elevation feedbacks to make ice margins more mobile under higher resolution; it is simpler for the ice sheet to advance as grid size $\Delta x \rightarrow 0$. Strategies involving adaptive grids (e.g., Price et al., 2008) or sub-grid tracking of the ice sheet margin are needed to give convergent behavior that is insensitive to model resolution.

8.3.2.4.3. Uncertainties associated with boundary conditions

H1_Geothermal heat flux

Pronounced subglacial topography with a very deep basal valley such as that of Jakobshavn Isbræ intensifies the local geothermal heat flux by as much as 100% (van der Veen et al., 2007), potentially introducing an important feedback between subglacial topography and fast flow. On a broader scale, significant variations in crustal thickness have been inferred from airborne gravity, with the thinnest crust underlying the Northeast Greenland Ice Stream and other major drainage routes (Braun et al., 2007). Where the crust is relatively thin, geothermal heat flow may be expected to be greater than the continental average (Leftwich et al., 2007). High geothermal heat flux in northern Greenland is also believed to play a role in providing the temperate, lubricated bed that enables fast flow in this ice stream (Fahnestock et al., 2001). The extent to which spatial variations in heat supplied to the glacier base affect the regional flow regime remains to be resolved (Greve, 2005). This could explain discrepancies in areas where basal meltwater has been detected but where models incorrectly predict the basal ice to be frozen to the bed.

H1_Climatic forcing

Ice sheet models require distributed (spatially-resolved) atmospheric fields to provide SMB fields. To examine climate change scenarios, temporal evolution of atmospheric conditions needs to be characterized. This appears as a boundary condition in ice sheet models, and is prescribed simplistically in most ice sheet model experiments. Uncertainties in spatial-temporal climate evolution may well dominate uncertainties in ice dynamics modeling for prediction of Greenland Ice Sheet response to climate change. This is discussed in more detail in Section 8.3.3. Still striving to present a correct present-day mass balance, the challenge of reproducing past and future mass balances is large.

In addition to atmospheric fields, ocean and sea-ice conditions directly impact on marine outlet glaciers (e.g., Holland et al., 2008; Section 8.2.3). Marine influences are parameterized through the

ocean heat flux that is prescribed as a basal boundary condition for floating ice; this is generally held fixed in current models, for example, a uniform flux of 2 W/m^2. Plus, ocean conditions can be incorporated in calving parameterizations, but this has rarely been done, in part because of unclear mechanistic links. Except for a small number of regional studies, ocean and ice sheet models are largely uncoupled and ocean variability is not considered in ice sheet modeling. Challenges in ice sheet modeling are revisited in the summary comments of Section 8.3.4, where the main sources of uncertainty for predicting Greenland Ice Sheet sensitivity to climate change are discussed.

8.3.3. Future climate and ice sheet scenarios: Response of the Greenland Ice Sheet to a warmer climate

8.3.3.1. Objectives of coupled ice sheet-climate modeling

Ice sheet models, as described in Section 8.3.2, can investigate the ice-dynamical response to idealized climate change scenarios, such as a uniform cooling or warming. However, patterns of change in climate fields (e.g., temperature, precipitation, radiation, winds) will not be uniform, as used in the models, over the area and elevation range of the Greenland Ice Sheet. Realistic climate forcing is required to provide more detail on ice sheet evolution in the coming decades and centuries. Ice sheet evolution over long timescales also introduces a number of feedbacks that need to be captured in future climate forecasts, such as changing elevation and albedo fields over the ice sheet. Coupled ice sheet-climate models are needed to capture these feedbacks.

The next section describes the coupling approaches adopted in ice sheet-climate modeling. Model predictions of Greenland Ice Sheet response to future warming are then summarized, followed by a discussion of some of the limitations of current models and priorities for future development.

8.3.3.2. Coupling approaches

8.3.3.2.1. One-way and two-way coupling

One-way and two-way approaches have both been adopted for coupling atmospheric climate and ice sheet models. In one-way coupling, output from the climate model provides input for the ice sheet model (e.g., SMB fields), but the ice sheet model output does not force the climate model. This allows existing climate change scenarios (e.g., the ensemble of IPCC climate model forecasts) to be applied to ice sheet simulations. To increase realism of the exchange, relationships between surface elevation, latitude, and surface conditions have been prescribed to allow climate forcings to change as the ice sheet surface evolves (Huybrechts et al., 1991). This includes elevation-temperature (lapse rate) feedbacks, precipitation rates that increase with increasing temperatures, roughly in accord with a Clausius-Clapeyron relationship, and a modification of local precipitation rates as a function of local surface slope (Ritz et al., 1997).

In two-way coupling, climate model output forces the ice sheet model, while output from the ice sheet model (e.g., ice topography, albedo fields) serves as input to the climate model. Two-way coupling is desirable as the simulation more realistically represents the physical system and the climate feedbacks associated with temporal evolution of the ice sheet. However, as two-way coupling is more challenging than one-way coupling and there are, generally, discrepancies of resolution between climate and ice sheet models, most ice sheet modeling experiments to date have used one-way coupling.

The model results in Figures 8.25, 8.26, 8.27, and 8.28 are examples of ice sheet simulations with one-way climate forcing. SMB is estimated from temperature and precipitation fields, using a degree-day methodology to estimate surface melt and the fraction of precipitation to fall as snow (Reeh, 1991). Present-day precipitation and temperature fields are taken as the baseline and temporal variability is prescribed as a perturbation based on ice core reconstructions, climate model forecasts, or simple scenarios (e.g., a step warming of 2 or 4 °C). Many Greenland Ice Sheet modeling studies have adopted this approach (Huybrechts et al., 1991, 2004; Ritz et al., 1997; Huybrechts and de Wolde, 1999; Greve, 2000; Alley et al., 2005; Gregory and Huybrechts, 2006).

While several parameterized ice-climate feedbacks are included in these numerical experiments, the one-way approach may omit key physical processes that affect the circulation of the atmosphere and ocean, which in turn would modify temperature and accumulation patterns on the ice sheet. A small number of groups have begun to interactively couple ice sheet models with climate models of varying degrees of complexity and resolution. Huybrechts et al. (2002), Fichefet et al. (2003), and Driesschaert et al. (2007) forced an ice sheet model with an AOGCM and examined how Greenland Ice Sheet retreat may affect oceanic circulation in the coming decades. These studies use one-way ice-sheet forcing from the atmospheric model. Other studies focus on ice sheet-atmosphere feedbacks (Gregory et al., 2004; Toniazzo et al., 2004), while a small number of studies explore fully-coupled ice sheet and AOGCMs (Ridley et al., 2005; Mikolajewicz et al., 2007b; Vizcaíno et al., 2008).

A number of climate-system and ice sheet feedbacks identified in coupled model studies are discussed in Section 8.3.3.3. Specific predictions of Greenland Ice Sheet retreat and sea-level rise in response to climate change are summarized in Section 8.3.3.4.

8.3.3.2.2. Climate downscaling

The experiments described above require atmospheric fields to be downscaled to the ice sheet model grid for mass balance modeling. This is due to the mismatch that exists between even high-resolution AOGCMs running at 1.25° (~150 km) grid resolution and ice sheet models (10 to 20 km). Grid downscaling is thus required and commonly takes the form of a simple interpolation.

This step requires a good understanding of altitude-SMB relationships in different climate regions of the ice sheet. Constant and spatially-uniform temperature lapse rates (linear temperature decreases with elevation) are usually used to downscale temperatures, often with different summer and annual lapse rates (Huybrechts et al., 1991). This does not account for the effects of inversions, katabatic-wind induced cooling at low elevations on the ice sheet, or surface energy balance effects that govern near-surface (*vs* free-air) temperatures, so a uniform lapse rate is a simplification. Precipitation is more difficult to downscale, as it can vary over small spatial scales, particularly in association with orographic forcing of precipitation at the ice sheet margin. Most modeling studies to date are based on perturbations to the present-day spatial pattern of precipitation in Greenland.

Under interactive coupling, the influence of significant changes in ice sheet topography and boundary conditions are better accounted for. For example, processes such as orographic precipitation and the energy balance influence of increasing areas of open water and ice-free land can be included as feedbacks in the model, rather than an assumption that present-day climate patterns will hold.

Because of the uncertainty associated with interpolation of meteorological fields, the downscaled climate is typically parameterized into positive-degree-day (PDD) input in order to simulate snow/ice melt; a full surface energy balance is not generally attempted. PDD approaches have proven reasonable for melt modeling in Greenland, but downscaling or interpolation of temperature or PDD fields does not ensure conservation of energy between the applied surface climatology and the climate model. There is a similar concern for downscaled precipitation fields. Conservation of mass with respect to larger-scale climate model precipitation and moisture fields is not straightforward and is often disregarded. Conservation of energy and mass warrant serious attention, particularly in two-way coupling experiments where there is a possibility of cumulative biases leading to 'drift' in transient simulations.

8.3.3.2.3. Mass balance estimation in coupled models

For the reasons outlined above, most Greenland Ice Sheet modeling studies to date estimate SMB from degree-day calculations as a function of climate (i.e., temperature and precipitation) perturbations. As was discussed in detail in Section 8.2.2, SMB fields have also been simulated directly from climate model re-analyses (Hanna et al., 2002, 2005, 2008), remote sensing studies (Mote, 2003), and regional climate modeling (Box et al., 2006; Fettweiss et al., 2007, 2008). Like the reconstructions by Hanna and co-workers, regional climate models make use of re-analyzed

climatology to provide the boundary conditions for atmospheric simulations. The main difference is the nature of accumulation and melt modeling in the two approaches; Hanna and co-workers use a degree-day approach, while regional climate models distribute precipitation based on atmospheric dynamics and estimate melt from a surface energy balance.

Table 8.2 gives a summary of simulated SMB terms from the historical period for several of these efforts. These studies provide detailed estimates of Greenland Ice Sheet mass balance through the second half of the 20th century, but future forecasts cannot be carried out with re-analyzed climatology. RCMs offer a potential solution to this, as they can be trained/calibrated on the historical period and forced by global-model fields for future simulations. This has yet to be attempted for Greenland.

Regional climate models have the inherent capacity to represent the physical processes of how snowfall and melt relate to the local and regional surface characteristics and energy budgets, and these models are likely to form the basis for the next generation of Greenland Ice Sheet predictions. However, climate models exhibit systematic biases, for example, land surface temperature too high or too low, in response to biases in model surface energy budget. These errors have been attributed largely to cloud-radiation bias, such as too much or too little downward longwave radiation. Terrain biases are another typical source of systematic bias, commonly arising from inaccurate representation of the terrain or from the need to smooth the surface topography at the 'steep' ice sheet margin to avoid atmospheric dynamical instabilities, for example, gravity waves. Through comparison with in situ observations, calibration procedures can be developed and uncertainties can be quantified (e.g., Box et al., 2004, 2006). However, a careful assessment of the stationarity of model biases is needed before these models can be applied with confidence to future forecasts.

8.3.3.2.4. Ice sheet-ocean coupling

Regional-scale coastal and sub-ice shelf oceanographic models are also needed to improve ice-climate coupling, something that has been explored on a regional scale in Antarctica, but has not been considered in Greenland. Other than an influence of global sea level on calving rates, ice-ocean feedbacks have not been examined on the scale of the whole ice sheet; ocean temperatures and heat fluxes, sea-ice conditions, and other potential controls of iceberg calving and melting at the ice-ocean interface are invariant in current models. This point is discussed further in Section 8.3.4.

8.3.3.3. Feedbacks in coupled models

Simulations with two-way coupling of the ice-ocean-atmosphere system reveal positive and negative feedbacks on ice sheet retreat in a warmer world. Huybrechts et al. (2002) and Fichefet et al. (2003) examined the coupled response of a 3-D thermo-mechanical model of the Greenland Ice Sheet and an AOGCM to future climate warming (IPCC B2 emissions scenario, mid-range greenhouse gas forcing). Simulated ice sheet melt is put into the ocean circulation model. Modeled melt rates are modest in the 21st century (global sea level rise of 4 cm in response to a warming of 4.5 °C over Greenland), but Fichefet et al. (2003) concluded that the freshwater addition to the ocean from ice sheet melt would contribute to a weakened Atlantic Meridional Overturning Circulation (AMOC) by the end of the 21st century; a feedback that lowers temperatures over East Greenland and partly counteracts warming in this region.

In a subsequent study, Driesschaert et al. (2007) concluded that Greenland Ice Sheet melting only significantly impacts on the AMOC in the 21st century under severe future warming scenarios. This is echoed in the results of Mikolajewicz et al. (2007b) and Vizcaíno et al. (2008), who used a coarse-resolution AOGCM interactively coupled with an ice-sheet model at 80-km resolution to explore millennial-scale future climate and ice sheet changes in Greenland. They also predicted that a weakened AMOC will reduce the degree of warming over southern Greenland (Figure 8.29), buffering Greenland Ice Sheet melt. However, changes in the hydrological cycle of the North Atlantic region are the main reason for this; Greenland Ice Sheet runoff is about 10% of the total freshwater forcing of the system (Mikolajewicz et al., 2007b).

Vizcaíno et al. (2008) predicted that a doubling or tripling of atmospheric CO_2 would result in 2 to 5 °C of summer warming in Greenland, with greater warming in northern Greenland (Figure 8.29). Simulations with 4 × CO_2 trigger a shutdown in North Atlantic deepwater formation, which creates a dramatic reversal in the projections and a cooling of up to 3 °C in southern Greenland.

Atmospheric feedbacks in response to Greenland meltback have been examined extensively through the productive collaboration of Philippe Huybrechts, Jonathan Gregory, and colleagues, using Huybrechts' ice sheet model and the Hadley Centre AOGCM. Toniazzo et al. (2004) examined the effect of removing the Greenland Ice Sheet on regional atmospheric circulation and climate. A lower (or absent) ice sheet reduces the persistent anticyclonic flow over the ice sheet, particularly in winter months, causing a net increase in precipitation in eastern and northern Greenland. Low elevation, reduced albedo, and the development of vegetation (e.g., boreal forest) cause greatly intensified summer warming and seasonal melting. No perennial snow is predicted. This prevents the ice sheet from re-growing in today's climate; it is an artifact of the Pleistocene, preserved by its own elevation and regional cooling influence.

Ridley et al. (2005) examined two-way coupling for a 4 × CO_2 climate. In this analysis, the ice sheet model output forces the climate model through changes in ice topography, surface albedo, and freshwater runoff. Ridley et al. (2005) concluded that long-term trends are primarily sensitive to topographic feedbacks (i.e., elevation-temperature effects) that are reasonably well captured in the one-way forcing parameterizations of, for example, Huybrechts et al. (1991). There are some additional feedback effects that appear to slow down ice loss. Warming of the ice-free region at the ice sheet margin, which intensifies under climate warming (and with ice sheet retreat) generates a convective circulation that cycles warm air to the interior of the ice sheet, while enhancing the intensity of katabatic return flows at the surface. This cools the lower elevations, reducing melt rates at the ice margin and helping to preserve the ice there.

Ice dynamics introduce another negative feedback in the models during early stages of the ice sheet retreat. As described by Huybrechts and de Wolde (1999) and Ridley et al. (2005), higher ablation at the margins and increased snowfall at high elevations combine to give greater balance gradients and steeper ice sheet slopes. This produces an increased ice flux which buffers drawdown at the ice sheet margin, maintaining thicker ice and reducing the elevation-temperature feedbacks. This reduces the initial rate of ice sheet retreat, although it is not sustainable as drawdown in the interior regions eventually causes an expansion of the ablation area, with elevation and albedo feedbacks that hasten ice sheet retreat.

8.3.3.4. Predicted Greenland Ice Sheet response to warming

Bearing in mind the missing ice-dynamical processes and ice sheet-climate coupling challenges, a number of insights are available from simulations to date. It is clear that the Greenland Ice Sheet is very sensitive to climate warming. In steady-state, a regional warming of more than 3 °C is likely to be sufficient to precipitate ice sheet retreat in southern Greenland (Huybrechts et al., 1991; Alley et al., 2005). This is illustrated in Figure 8.28; the southern dome retreats by 3000 AD in these simulations at CO_2 levels between 550 and 750 ppm, which correspond to a summer warming of 4 to 5 °C in Greenland.

Based on an ensemble of models, including high-resolution (~1°) climate model representations of temperature and precipitation patterns over the ice sheet, Gregory and Huybrechts (2006) concluded that an average annual regional warming of 4.5 °C would push the Greenland Ice Sheet into a negative SMB. The threshold temperature is higher than in earlier studies (e.g., 2.7 °C in Huybrechts et al., 1991; 3 °C in Gregory et al., 2004), in part because summer warming over Greenland, ΔT_S, is forecast to be less than the mean annual warming, ΔT_A: Gregory and Huybrechts (2006) reported an average value $\Delta T_S / \Delta T_A = 0.81$. Summer temperature is the critical parameter for Greenland's mass balance, as this is the main driver of annual surface melt and runoff.

A warming of 4.5 °C in Greenland corresponds to an average global warming of 3.1 °C in the ensemble of models explored by Gregory and Huybrechts (2006). The relationship is plotted in Figure

8.30, along with estimates of rates of sea level rise due to SMB losses as a function of mean annual warming. There is an average polar amplification factor of ~1.5 over Greenland, less than the average value for the Arctic. Part of the moderate polar amplification factor here may be due to the cooling influence of weakened deepwater formation in the Nordic Seas that is predicted by many AOGCMs in the coming decades (see for example Figure 8.29). This is addressed in more detail in Section 8.4.1.

The critical warming value of 4 to 5 °C for Greenland to go into a state of negative SMB is often quoted as a threshold for 'irreversible' ice sheet decline. Dynamical ice losses would place the Greenland Ice Sheet into a negative total mass balance well before this SMB threshold was met – this threshold appears to have been reached in the 1990s and 2000s with a regional warming of close to 1 °C (Section 8.2.4). However, the temperature threshold for SMB to become negative is considered to be terminal because ice sheet thinning would provide a positive feedback to ice sheet decay. If this proceeds too far, ice sheet drawdown would make it difficult to reverse the decline, even with temperature stabilization. However, the timescale for such a decline is long – many centuries to millennia – so it is difficult to judge irreversibility on this timescale.

Charbit et al. (2008) argued that 3000 Gt of cumulative CO_2 emissions could lead past this threshold for an irreversible Greenland Ice Sheet retreat, such that anthropogenic activity in the 21st century could commit the world to several centuries of sea level rise. Ridley et al. (2009) examined this question through coupled modeling with climate restored to its pre-industrial state for several different Greenland Ice Sheet configurations. They argued for a potential two-stage destabilization of the ice sheet in Greenland, with reductions to 80–90% of the current ice sheet volume inducing ice loss in southern Greenland and an irreversible sea level rise of about 1.3 m over several centuries. If the ice sheet retreats further before climate is stabilized (i.e., under greater or sustained warming), such that the ice sheet falls below half of its current volume, Ridley et al. (2009) predicted an irreversible sea level rise of about 5 m.

Such a decline is consistent with reconstructions of a reduced Greenland Ice Sheet in the last interglacial period (Cuffey and Marshall, 2000), in response to orbitally induced spring and summer warming at high northern latitudes. Other ice sheet modeling studies predict a less severe ice sheet decline at this time (Huybrechts et al., 1991), due to different assumptions about the ice core isotope-temperature relationship. Miller et al. (2006) and Otto-Bliesner et al. (2006) estimated a summer warming of about 5 °C near Greenland at the peak of the Eemian warming. Ice sheet models predict an almost complete collapse of the ice sheet if warming of this magnitude persists for several thousand years, but the modeled retreat takes several millennia – changes over the first few centuries are modest.

For future warming, the extent of modeled Greenland Ice Sheet retreat by 2100 is primarily a function of the change in SMB, hence the climate scenario. Ice dynamics plays only a minor role over this time scale in the current models, for two main reasons: elevation-induced SMB and ice-dynamic feedbacks are minor over a one-century timescale, and ice dynamics models lack a direct connection with climate (e.g., ocean or surface-meltwater forcing). The latter is consistent with ice-sheet model simulations of the past century; there is little to no interannual variability in modeled ice dynamics. This is partly because ice dynamics in the current generation of ice sheet models is sensitive only to local gravitational driving stress (see Section 8.3.2), which changes slowly.

Table 8.5 provides a summary of future sea-level rise forecasts from model studies of Greenland Ice Sheet response to climate change. These studies are difficult to compare directly because they make different assumptions about future climate change, the climate downscaling strategy and SMB models differ in each study, and results are not always documented for 2100. These results all include ice sheet dynamical feedbacks.

Table 8.5. Model projections of future sea level rise (SLR) associated with Greenland Ice Sheet retreat. Results are from ice sheet modeling studies with one-way climate forcing introduced as a perturbation to present-day temperature and precipitation fields, unless noted otherwise.

Year	SLR,	Souce	Climate treatment

	cm		
2100	4.8	Huybrechts et al., 1991	Warming anomaly increasing to +4.2 °C by 2100
2100	3.9	Huybrechts and de Wolde, 1999	Radiative forcing for stabilization at 2 × CO_2
2100	4.0	Huybrechts et al., 2002	IPCC B2 emissions scenario, single AOGCM
2100	2-7	Huybrechts et al., 2004	IPCC IS92a emissions scenario, ensemble of AOGCMs
2100	1-9	Alley et al., 2005	Range of AOGCM future climate scenarios
2100	10.4	van de Wal and Oerlemans, 1997	IS92a radiative forcing, energy balance model
2100	6-14	van de Wal et al., 2001	As above but a range of IS92 radiative forcings
2070	1.3	van de Wal et al., 2001	2 × CO_2, AOGCM forcing of energy balance model
3000	88-317	Huybrechts and de Wolde, 1999	Radiative forcing, stabilization at 2× and 4 × CO_2
3000	72-240	Greve, 2000	Temperature anomalies of +3 and +6 °C
3000	25	Vizcaíno et al., 2008	2 × CO_2, fully coupled ice model and AOGCM
2600	98	Vizcaíno et al., 2010	2 × CO_2, fully coupled ice model and AOGCM

Under CO_2 stabilization at concentrations of 2× pre-industrial, equivalent to the IPCC IS92a climate change scenario, Huybrechts and de Wolde (1999) simulated a contribution to sea level rise from the Greenland Ice Sheet of 3.9 cm by 2100. Equivalent simulations for 4× and 8× pre-industrial CO_2 concentrations give ice sheet retreat equivalent to 10.6 and 20.8 cm of sea level rise. Using a more sophisticated climate treatment with the IS92a scenario, Huybrechts et al. (2004) estimated an average 21st century rate of sea level rise of 0.45 to 0.49 mm/y from an ensemble of climate models: a net sea level rise of 4.5 to 4.9 cm by 2100 (Figure 8.31). The range of model estimates for 2100 from Huybrechts et al. (2004) is 2 to 7 cm; these estimates are all for the IS92a emissions scenario, so this represents the variation in climate models. Alley et al. (2005) gave a similar estimate for 2100, 5 ± 4 cm, for a fuller range of emissions scenarios.

Greenland Ice Sheet retreat has also been studied by van de Wal and Oerlemans (1997), with a simpler (2-D, vertically-integrated) ice sheet model but using an energy balance model to calculate surface melt. They estimated 10.4 cm of sea level rise from Greenland between 1990 and 2100 due to changes in SMB under the IS92a scenario. Depending on assumptions about basal sliding and the initial state of the ice sheet, van de Wal and Oerlemans (1997) estimated a range from 5 to 16 cm of sea level rise. In a follow-up study, van de Wal et al. (2001) examined the impact of different emissions scenarios and reported a similar range of sea-level rise forecasts, from 6 to 14 cm by 2100. Sea level rise accelerates as the century progresses. The reference scenario, IS92a (10.4 cm), corresponds with 4.1 cm of sea level rise by 2070. With climate fields derived from an AOGCM for a similar scenario (2 × CO_2), van de Wal et al. (2001) estimated only 1.3 cm of sea level rise by 2070, which can be extrapolated to 3.3 cm by 2100. Increased snow accumulation in the AOGCM explains the lower total in this study, compared to the energy balance model results driven by the IS92a reference scenario.

The analysis by Gregory and Huybrechts (2006) is the most detailed examination to date of the expected future contributions to sea level rise from warming-driven changes in SMB. Figure 8.32 plots the phase space of simulated sea level rise as a function of change in temperature and precipitation. Decadal changes in temperature and precipitation from a suite of AOGCM simulations for the 21st century indicate projected temperature increases of up to 8 °C and precipitation increases of up to 50%. The net effect on SMB is negative in the models; higher melt rates beat increases in accumulation. The corresponding rate of sea level rise due to declining SMB ranges from 0 to 2 mm/y for warmings of 0 to 8 °C. Examining Greenland Ice Sheet sensitivity to changes in temperature and precipitation, along with an ensemble of AOGCM forecasts for the full spectrum of IPCC AR4 emissions scenarios, Meehl et al. (2007) reported a Greenland Ice Sheet contribution to global sea level of 1 to 12 cm by 2100.

Based on this collection of model studies, the direct SMB effect of expected warming in Greenland is likely to be a sea level rise of 5 to 10 cm by 2100, corresponding to 200 to 400 Gt/y loss of mass in

average over the period to 2100. Within the range of anticipated climate change, ice volume losses in excess of 1 m of sea-level equivalent from the Greenland Ice Sheet require several centuries in all published modeling studies to date. The complete demise of the ice sheet requires 3000 years or more under most future climate change scenarios. The most severe climate warming simulated by Greve (2000), that is, 12 °C over Greenland, precipitates a full collapse (and a 7 m sea level rise) in 1000 years. Predictions by Greve (2000) for more moderate warming scenarios of 3 and 6 °C give a sea level rise of 0.72 m and 2.4 m after 1000 years. Contributions by 2100 are about 5% of this. Huybrechts and de Wolde (1999) found a similar rate of extreme ice sheet retreat in their most severe climate change scenario, 8 × CO_2, with an average rate of sea level rise of 60 cm per century and about 80% of the Greenland Ice Sheet melting within 1000 years under these conditions. Rates of ice loss increase non-linearly with time in these simulations, as elevation-climate feedbacks strengthen.

While the forecasts are reasonably well bounded for this century, models have a conservative bias for ice sheet response to climate change, due to missing processes in the glaciological models (see Section 8.3.2) and a number of missing climatic feedbacks. There is no sensitivity to changing conditions at the ocean interface in current models, that is, no capacity for outlet glacier acceleration.

Pfeffer et al. (2008) probed this by considering an extreme case where all of the marine-based outlet glaciers of the Greenland Ice Sheet double their discharge to the ocean. By taking the cross-sectional area of all marine outlet glaciers and assuming a constant ice thickness at these flux gates, Pfeffer et al. (2008) calculated that a two-fold increase in outlet glacier velocity would contribute 9.3 cm to sea level rise by 2100. This is an extreme case: all of the outlet glaciers would need to double their velocity immediately and remain at this higher rate of flow for the rest of the century, without significant thinning. This, therefore, provides a plausible upper bound on ice dynamical discharge in Greenland.

Pfeffer et al. (2008) added this term to the sea level rise associated with SMB losses to give an upper bound for Greenland mass loss. However, the Greenland Ice Sheet's 'baseline state' and current estimates of future sea level rise from Greenland (e.g., Huybrechts and de Wolde, 1999; Greve, 2000; Gregory et al., 2004; Huybrechts et al., 2004) include contributions from dynamical ice losses (i.e., iceberg calving). Current models have negligible 21st century variability in this term and muted sensitivity to climatic forcing, relative to recent observations, but there is still a background level of ablation from this mechanism. Hence, it might be appropriate to consider only the extra iceberg discharge associated with a two-fold speed-up of marine outlet glaciers in placing an upper bound on Greenland's contribution to sea level rise. Using the fluxes calculated by Pfeffer et al. (2008), this gives 4.7 cm of sea level rise by 2100. If this is based, instead, on the estimated baseline ice discharge from Greenland in the early 1990s, 350 km^3/y (Table 8.3), the additional discharge from a doubling of the ice flux would give a 21st century sea level contribution of 8.8 cm. This can be added to the expected SMB-driven losses of up to 10 cm to give an upper bound of 19 cm of sea level rise from the Greenland Ice Sheet by 2100 corresponding to an average mass loss of 800 Gt/y in the period to 2100. This is not a true upper bound. The value could be higher, if climate change during this century is more severe than expected in Greenland, or as a result of additional climate feedbacks that are lacking in the current models.

8.3.4. Summary and outlook

The complexities of climate and ice sheet models are such that much work is still needed to refine model physics, improve model resolution, and assimilate observational data to better constrain and test simulations of the meteorology, mass balance, and dynamics of the Greenland Ice Sheet. Great strides have been made in coupling of ice sheet and climate models in recent years. RCMs now offer the possibility of grid-matching and a full surface energy balance calculation, to help reduce uncertainties associated with climate downscaling and some of the simplistic assumptions in current parameterizations of SMB.

A number of climate forcings and ice sheet-climate processes are still missing in current models, particularly at the ice-ocean interface. This is pressing in light of evidence that oceanic forcing is responsible for some of the recent, rapid changes in the Greenland Ice Sheet mass balance and ice

dynamics (Holland et al., 2008). Most fast-flow processes in ice sheets are not well represented in current models, and potential climatic excitation of fast flow, through either the ocean or atmosphere, has not been explored. This means that the parameter space for Greenland Ice Sheet response to climate change is not fully explored; models cannot give reliable bounds on this.

The kinematic approach of Pfeffer et al. (2008) probably offers the best available upper bound on the ice-dynamical contributions to mass balance losses in Greenland. For a proper dynamical simulation of these effects, the ice-dynamical processes discussed in detail in Section 8.2.3 need to be captured by predictive models. This is important to improved estimation of the total ice sheet mass balance, given the potential for large-scale climate forcing of ice-dynamical ablation mechanisms. At present, ice sheet models are not sensitive to surface meltwater production or to ocean and sea-ice conditions, all of which have begun to change rapidly in the Arctic.

Further development of AOGCMs and regional equivalents is also needed. Little besides what has been reviewed here has been done to couple climate and ice dynamical models. Thus, more coupling experiments represent an obvious area of future work, particularly with respect to regional patterns of climate downscaling and atmospheric/mass balance feedbacks of changing ice sheet geometry.

Other climatic feedbacks on ice sheet mass balance are not fully explored, such as the impact of decreasing ice sheet albedo as the ablation zone of the ice sheet expands spatially and seasonally. Degree-day melt models typically use two melt factors, one for snow and one for ice, rather than the continuum of behavior that is evident in nature. The positive feedbacks of a wetter, lower-albedo snow surface may not be being captured in present model forecasts. More physically-based melt models (e.g., from energy balance calculations on high-resolution grids) will provide improved estimates of the threshold temperature at which the Greenland Ice Sheet switches into a state of negative SMB.

These limitations make it difficult to project the extent of sea level rise that can be expected from the Greenland Ice Sheet in this century and beyond. Current models suggest 5 to 10 cm by 2100, but conservative biases in the models mean that it may be in excess of this. If climate warming excites systematic, widespread acceleration of the outlet glaciers in Greenland, the ice sheet could contribute as much as 19 cm to global sea level rise this century. Extreme warming scenarios (e.g., 10 °C) are needed to extract more ice than this from Greenland by 2100, although long-term ice sheet losses may result from instabilities that are triggered in the coming decades.

The models are still in early stages of coupling and several key processes and feedbacks are absent or poorly parameterized, which compromises the ability of the models to simulate interannual variability in ice velocities (hence, ice discharge to the oceans), as well as some potentially important aspects of ice sheet sensitivity to climate forcing. Some of the key processes that need to be better understood and included in the models include:

- subglacial processes governing basal flow
- effects of surface meltwater on basal flow
- numerical representation of ice stream and outlet glacier dynamics
- oceanic influences on ice flow and ice-marginal ablation
- surface energy balance feedbacks as the ice sheet wanes (e.g., declining albedo).

These missing processes introduce a likely conservative bias in model studies to date, such that they may systematically underestimate Greenland Ice Sheet sensitivity to climate change. Recent observations and simplified (e.g., conceptual) model studies indicate that increased ice discharge associated with a warmer climate could effectively double the ice sheet sensitivity to climate change, with ice dynamics contributing 5 to 9 cm to sea level rise by 2100. This is on top of the SMB-driven ice sheet losses of 5 to 10 cm, giving a total contribution to sea level rise of 10 to 19 cm.

8.4. Impacts caused by changes in the Greenland Ice Sheet

- The mass loss from the Greenland Ice Sheet will increase and cause a rise in sea level that will impact globally, especially the vast population living close to the [coast / sea level].

- The increasing freshwater input to the fjords and ocean surrounding Greenland will change salinity and impact on the marine food web and thereby the fisheries.

- Freshwater input to the ocean will affect the thermohaline ocean circulation. But as the component from the Greenland Ice Sheet is less than 10% of the total freshwater input the effect of the increased melt is assumed to be small.

8.4.1. Ocean circulation

Freshwater from the Greenland Ice Sheet melt as well as iceberg calving into the sea have various effects on the waters of the Arctic region (i.e., Arctic Ocean with continental shelves, Nordic Seas, and Labrador Sea) as well as the World Ocean. The Greenland Ice Sheet runoff dilutes the marine surface water and reduces its salinity and density, thereby impacting global sea level rise and two types of ocean circulation: thermohaline circulation and estuarine circulation. All are important for the marine climate of the Arctic region although they are affected in different ways.

8.4.1.1. Sea level change

8.4.1.1.1. Global sea level change

The Greenland Ice Sheet represents, if entirely melted, a global average eustatic sea level rise of about 7 m. In a warmer climate with more precipitation, accumulation, ablation and ice discharge will increase but model studies suggest that the net balance of these factors will be toward a diminishing ice sheet (Gregory and Huybrechts, 2006). Depending on the future temperature evolution, the potential disappearance of the Greenland Ice Sheet or its transformation to a much reduced inland ice cap, will take of the order of millennia (Gregory et al., 2004).

The contribution to global sea level rise from the Greenland Ice Sheet has increased during the past decades (Cazenave et al., 2008) and global sea level has been observed to rise at a faster rate than projected by Meehl et al. (2007). Motivated by this apparent under-estimation of sea level rise from dynamical modeling, Rahmstorf (2007) suggested an alternative semi-empirical approach where a statistical model is fitted to historical time series of global sea level and temperature. By then driving the statistical model with temperature scenarios Rahmstorf obtained total global sea level changes of 0.5 to 1.4 m for the end of the 21st century; larger than by Meehl et al. (2007). This work has, however, been criticized by several authors (Holgate et al., 2007; Schmith et al., 2007; Rahmstorf, 2008; von Storch et al., 2008). An improved semi-empirical model by Grinsted et al. (2009) estimated the likely sea level rise to be in the range 0.7 to 1.6 m at the end of the 21st century.

8.4.1.1.2. Regional sea level change

The increased release of freshwater from the Greenland Ice Sheet will, together with the expected increase in net precipitation over the North Atlantic area, cause changes in the salinity and temperature distribution (Curry et al., 2003), which, through a dynamic response, will cause regional changes in sea surface height. This dynamic response can be either thermohaline in nature, such as weakening of the AMOC, or an adjustment of gyre circulations due to changes in the density distribution. Moving mass from the ice sheet to the ocean causes changes in the gravity field (Mitrovica et al., 2001; Milne et al., 2009). This has remained unnoticed to date, but causes noticeable regional differences in sea level rise. Diminishing the Greenland Ice Sheet is expected to produce the largest sea level increase in the South Atlantic and the North Pacific and a decrease in the North Atlantic and around Greenland. This does not take vertical isostatic land lift into account, which takes place on a longer time scale.

A quantitative assessment of the effect of freshwater release from the Greenland Ice Sheet does not exist at present. Landerer et al. (2007) analyzed a standard IPCC scenario run (i.e., without melting from the Greenland Ice Sheet) and discovered regional signals, most prominent a sea level increase of 0.3 m in the Arctic Ocean and a minor increase (i.e., up to 0.1 m) along the Greenland coast. In the center of the subpolar gyre, Landerer et al. (2007) found a decrease of the order of 0.1 m, which could be due either to changed density distribution or to changed wind forcing, or both.

8.4.1.2. North Atlantic circulation – deep convection and climate impact

Through thermohaline ventilation, the Arctic region is a significant source to the North Atlantic Deep Water as part of the global thermohaline circulation. Therefore, increased runoff from Greenland has the potential to affect the circulation of the Arctic region and the global ocean.

Off Greenland (Figure 8.33), the East Greenland Current flows southward along the shelf towards the Atlantic (Holfort et al., 2008). Together with the outflow through the Canadian Arctic Archipelago, this comprises a low-salinity surface outflow from the Arctic (Melling et al., 2008). The inflow, to compensate for the estuarine outflow, consists of warm and saline Atlantic water concentrated to the eastern part of the North Atlantic and which cannot be distinguished from the direct wind- and thermohaline-driven surface inflow to the Nordic Seas (Hansen and Østerhus, 2000). This inflow enters the Arctic in three distinct branches: a branch to the west of Iceland, a branch between Iceland and the Faroe Islands, and a branch between the Faroe Islands and Shetland (Østerhus et al., 2005; Hansen et al., 2008).

Through heat loss to the atmosphere, the inflow is cooled as it flows northward, and this increases its density. Brine rejection from freezing of seawater also has this effect whereas freshwater input decreases density. The overall effect is to separate the inflow into an upper low-density water mass and a sub-surface high-density water mass that is then returned to the North Atlantic.

The low-density water mass is returned as surface outflow in the East Greenland Current and through the Canadian Arctic Archipelago, which also carries most of the Bering Strait inflow of Pacific water to the Arctic Ocean (Rudels, 1989). The high-density water mass returns to the North Atlantic through several current branches, termed 'overflows' (Saunders, 2001), and passes through a number of deep channels across the Greenland-Scotland Ridge.

In terms of volume transport, the overflows export roughly twice as much water to the North Atlantic as the surface outflows of the East Greenland Current and the Canadian Arctic Archipelago. The formation of overflow water by thermohaline ventilation involves two steps: density increase and sinking. For the density to increase substantially the water has to be in contact with the atmosphere and this occurs by heat loss in the western part of the Nordic Seas and in the Barents Sea (Simonsen and Haugan, 1996). The sinking may occur through open ocean convection (Marshall and Schott, 1999) in the Greenland Sea and Iceland Sea, but this is only one of the mechanisms (Hansen and Østerhus, 2000) and convection in the Greenland Sea is not likely to be the main source of overflow water (Eldevik et al., 2009).

The overflow water is cold, much of it at temperatures below 0 °C, and is relatively saline compared to other cold water masses in the ocean. Thus, it is one of the densest water masses found anywhere in the World Ocean but, immediately after crossing the Greenland-Scotland Ridge, it mixes strongly with, and so entrains, ambient upper ocean water masses (Dickson and Brown, 1994; Østerhus et al., 2008). The resulting water mass is warmer and less dense, but also has a larger volume transport. Most of this water flows into the Labrador Sea where it contributes to the North Atlantic Deep Water together with the Labrador Sea Water formed locally by convection (Lazier, 1980; Yashayaev et al., 2008). North Atlantic Deep Water is thus from three sources: overflow water, Atlantic water entrained into the overflows, and Labrador Sea Water. Formation rates for some of these are debated (Haine et al., 2008), but there seems to be general agreement that they are of similar magnitudes.

North Atlantic Deep Water is carried south by the deep branch of the AMOC and is the main deep water source of the World Ocean, complemented by Antarctic Bottom Water formed by dense outflows from the Antarctic shelf regions. The AMOC is associated with heat transport on inter-hemispheric scales toward the ventilation areas where heat is released to the atmosphere. This transport of heat by the ocean is modest compared to the atmospheric heat transport (Trenberth and Caron, 2001) and its importance for climate has been challenged (Seager et al., 2002).

However, convincing evidence suggests that, through climate interactions, this ocean transport indeed plays a disproportionate role for fundamentals of the global climate (Rhines and Häkkinen, 2003; Rhines et al., 2008). This also finds support from tailored 'what if' climate model experiments, where artificial shutdown of the North Atlantic Deep Water production has been shown to strongly cool the regional climate of the Nordic Seas and Barents Sea but with cooling evident over the entire Northern Hemisphere (e.g., Vellinga and Wood, 2002; Sutton and Hodson, 2005). Apparently, this has occurred in other climatic periods, for example, during the deglaciation period where reduced thermohaline ventilation was likely to have been triggered by increased extreme freshwater supply from the melting glaciers to the ventilation areas (Tarasov and Peltier, 2005). Changes in the AMOC also have indirect impacts on the global climate. The formation of deep water and its subsequent spreading is crucial for maintaining not only the density but also the chemical structure and the biological conditions characterizing the World Ocean (Schmittner, 2005) with feedbacks also on atmospheric chemistry and climate.

A number of confounding factors, however, make quantifying the impacts of increasing runoff from the Greenland Ice Sheet difficult. First, runoff is not the only freshwater source in the Arctic region. In the present-day climate, total Greenland Ice Sheet runoff (0.018 Sverdrup[1]; Dickson et al., 2007) is almost an order of magnitude smaller than the estimated freshwater (including sea ice) transport southward through Fram Strait (0.15 Sv; Carmack et al., 2008), which flows through the areas into which the Greenland Ice Sheet runoff is released. It is also much smaller than the freshwater component of the Canadian Arctic Archipelago outflow (Carmack et al., 2008).

Various changes in the freshwater cycle of high latitude lands and oceans have been reported for the period since the 1950s (Peterson et al., 2006). Changes have been widespread and perhaps dramatic in the case of the Great Salinity Anomaly of the 1970s but, by attribution, anomalous glacial melt has played a minor role, contributing only about 5% to the observed ocean freshwater storage in the North Atlantic and Arctic region. In the future, the Greenland Ice Sheet runoff may well, in relative terms, increase more rapidly than the other sources but even in the most extreme case, the projected peak rate of melting estimated from a coupled climate-ice sheet model (0.06 Sv; Ridley et al., 2005) remains considerably smaller than the contribution from other sources today (Dickson et al., 2007). In idealized anthropogenic climate change simulations, absolute changes in atmospheric moisture transport are also found to exceed changes in runoff (Mikolajewicz et al., 2007a; Vizcaíno et al., 2008), which tends to downplay the isolated role of the Greenland Ice Sheet runoff on ocean circulation.

Second, the effects of increased freshwater supply depend on other factors. The intensity of the estuarine circulation is determined mainly by the rate of mixing between freshwater and seawater thus depending primarily on stratification and winds (Stigebrandt, 2000). For the thermohaline ventilation, the amount of freshwater that can reach an area of deep ventilation is a crucial factor. Most of the Greenland Ice Sheet runoff should be able to reach the ventilation areas in the Labrador Sea, but only runoff from the east coast of Greenland, north of Denmark Strait, can directly affect the offshore regions of the Nordic Seas where overflow waters are produced.

In the present-day climate, even this runoff seems to affect these areas only weakly because most of it is carried south through Denmark Strait by the East Greenland Current (Dickson et al., 2007; Holfort et al., 2008), depending on the wind conditions over the East Greenland Current (Stigebrandt, 2000), which may well change. Although part of the runoff from the ice sheet may initially bypass areas of ventilation, it is recirculated in the sub-polar North Atlantic and will, on decadal scales, dilute the entire sub-polar North Atlantic and spread to the Nordic Seas (Curry et al., 2003; Curry and Mauritzen, 2005; Peterson et al., 2006). The recent freshening has indeed been traced into the deep overflows from the Nordic Seas (Dickson et al., 2002).

In anthropogenic climate change projections, the effect on the AMOC of increased freshwater supply to the ventilation areas in response to an intensified atmospheric hydrological cycle (Cubasch et al., 2001) may also be partly or completely compensated for by increased salinity of the Atlantic inflow.

[1] A Sverdrop = 1 Sv = 10^6 m^3/s.

With enhanced evaporation at low latitudes, the source waters in the sub-tropical North Atlantic are becoming more saline (Curry et al., 2003) and this effect has been shown to stabilize the AMOC (Latif et al., 2000). The pathways of subtropical waters toward the ventilation areas, however, are dependent on the circulation of the subpolar gyre (Hátún et al., 2005), which again depends on Labrador Sea convection (Häkkinen and Rhines, 2004). No such direct compensation, however, will be in effect considering an increased Greenland Ice Sheet runoff resulting from a negative mass balance of the ice sheet, or anomalous freshwater outflow from the Arctic Ocean due to retreating Arctic sea-ice cover.

Although these freshwater fluxes will certainly add to AMOC weakening under anthropogenic climate change, this effect, including changes due to intensification of the hydrological cycle, will possibly be masked by the direct effects of atmospheric warming on the AMOC. In a comparison of 11 different climate models, Gregory et al. (2005) found weakening of the AMOC in idealized global warming experiments to be induced mainly by changes in surface heat flux and to a lesser extent by changes in freshwater fluxes.

In the IPCC AR4, Working Group 1 reviewed the status of climate model projections for the AMOC and found it 'very likely' that the strength of the AMOC will decrease in the 21st century (Meehl et al., 2007). The working group did not expect the AMOC weakening to lead to cooling over Europe and considered an abrupt transition during the course of the 21st century to be 'very unlikely'. However, none of the model studies above take the effects of ice sheets into account.

Very few studies exist where GCMs have been coupled to dynamic ice sheet models (Fichefet et al., 2003; Ridley et al., 2005; Winguth et al., 2005; Mikolajewicz et al., 2007a,b; Vizcaíno et al., 2008). It may also be argued that only in the recent study by Mikolajewicz et al. (2007a) was the coupling complete. This model alone features a realistic climate, allowing reasonable simulation of ice sheets. Nevertheless, the models yield similar meltwater input rates and most, although not all, reveal only a moderate effect of enhanced Greenland Ice Sheet runoff on the AMOC. Thus, it would be premature to conclude on the effects of the Greenland Ice Sheet runoff on the thermohaline circulation on the basis of these few studies.

Whereas fully coupled models will be required to convincingly project future melt rates of the Greenland Ice Sheet and project future global sea level rise, the freshwater impact on the AMOC by other and probably more dominating sources may be studied independently. This understanding is crucial for assessing the stability of the AMOC under anthropogenic climate change where a large spread in projected weakening for the 21st century among available models (Gregory et al., 2005) indicates a significant model uncertainty. In particular, differences in the projected changes in the hydrological cycle explain differences in the projections of the AMOC (Cubasch et al., 2001; Vellinga et al., 2008). This is one of the key questions to be addressed in the FP7-funded research project THOR (ThermoHaline Overturning - at Risk?) initiated in 2008.

It has been debated whether the projected AMOC decrease has already been initiated. Bryden et al. (2005) suggested that the meridional overturning circulation decreased by 30% between 1950 and 2000 at 24 °N in the Atlantic Ocean with most of the decrease being due to declining overflows. This supports the suggestion of a similar, but smaller, decrease in the overflow transport through the Faroe Bank Channel (Hansen et al., 2001).

Direct measurements have demonstrated a high stability of this flow between 1995 and 2005 (Hansen and Østerhus, 2007) and Olsen et al. (2008) found no weakening of either this overflow branch or the total overflow transport from 1948 to 2005 in a study combining measurements with an ocean model. Concerns have also been raised about the methodology of the Bryden et al. (2005) study (Baehr et al., 2007; Cunningham et al., 2007; Wunsch, 2008).

Conversely, there is evidence in the Labrador Sea of a weakened convection since the mid-1990s, but the formation rate of Labrador Sea Water exhibits large variability (Haine et al., 2008) and attribution of this weakening to anthropogenic causes would be premature and, on the present observational basis (Peterson et al., 2006), it seems unlikely that enhanced Greenland Ice Sheet runoff has played a prominent role.

8.4.1.3. Sea ice

Sea-ice formation depends on the stability of the surface layer of the ocean. Changes to the surface layer incurred by increased meltwater runoff from the Greenland Ice Sheet in a warmer climate could therefore potentially influence sea-ice production.

8.4.1.3.1. Freshwater flux

An increased freshwater flux may lead to increased stability of the upper ocean, which could precondition greater sea-ice production. On the other hand, ice cover of the marginal seas around Greenland has reduced during the present global warming period (satellite data from the period 1978 to 2008). The total effect of higher annual mean air temperatures and increased freshwater flux is not well predicted.

In the Greenland Sea, the role of local sea-ice production and advection was shown to potentially precondition the ocean for deep convection (Visbeck et al., 1995; Tsubal and Coon, 2001). Brine rejection during sea-ice formation destabilizes the ocean and advection of the ice out of the area removes excess freshwater. After a period of ice formation and advection, the density gradient will be eroded and further cooling can generate deeper convection. With sufficient cooling, this mechanism can remove excess freshwater from the surface of the Greenland Sea and counteract an increased freshwater flux from Greenland in relation to deep water formation. However, as the mechanism is reliant on subsequent cooling during winter, it is not clear if it will prevail in a warmer climate.

Few examples of coupled global climate model predictions of the response of sea ice to an increased freshwater flux from the Greenland Ice Sheet have been published. Fichefet et al. (2003) showed indications of a possible feedback mechanism where a breakdown of the thermohaline circulation leads to a significant cooling over eastern Greenland and northern North Atlantic waters, amplified by the sea-ice albedo feedback related to an expansion of the sea-ice cover in the Greenland-Iceland-Norwegian seas. The cooling subsequently reduces the freshwater flux from Greenland.

H1_Feedback of less ice on the state of ice shelves

The reduced summer ice cover at the Arctic Ocean margins (e.g., Greenland Sea and Barents Sea) exposes the coastline of these regions to the influence of ocean swells. This may cause increased coastal erosion but may also directly influence the stability of floating glaciers, such as Storstrømmen and 79-fjorden, in northeastern Greenland.

Reeh et al. (2001) studied the interaction between sea ice and glacier, in the case of the floating tongue of the 79-fjorden glacier, northeastern Greenland (79°30' N, 22° W). The authors used information from glaciological and geological studies, expedition reports, aerial photographs and satellite imagery to document the position of the glacier front and fast-ice conditions on millennial to decadal time scales. The study indicates that stability of the floating glacier margin is dependent on the presence of a protecting fast-ice cover in front of the glacier. In periods with a permanent fast-ice cover, no iceberg calving occurs, but after fast-ice break-up the glacier responds with a large ice discharge activity, whereby several years worth of accumulated glacier ice flux suddenly breaks away.

Climate-induced changes in sea-ice conditions in the Arctic Ocean with seasonal break-up of the nearshore fast ice could lead to the disintegration of floating glaciers. The present dominant mass loss by bottom melting would then be largely overtaken by grounding line calving of icebergs and the local influx of freshwater from north Greenland glaciers to the sea would be reduced and local iceberg production would increase.

8.4.1.4. Greenland coastal currents, including fjords, circulation and stratification

The ocean currents around Greenland are part of the cyclonic sub-polar gyre circulation of the North Atlantic and the Arctic region (Figure 8.33). The surface waters in southeastern and western

Greenland are dominated by two very different water masses: warm and saline Irminger Water, a side branch of the North Atlantic Current, and cold and low-saline Polar Water originating from the Arctic Ocean.

These water masses meet in the northern Irminger Basin and in the Denmark Strait. The main branch of the Irminger Current turns west toward East Greenland where it meets the Polar Water, which is transported southward along East Greenland within the East Greenland Current. The strength of these two currents determines the hydrographic conditions around southeastern and West Greenland. The two water masses flow side by side forming large meanders at the front where intensive mixing takes place. Polar Water is found at the surface on the continental shelf whereas Irminger Water is found over the continental slope and partly below the Polar Water on the deeper parts of the continental shelf. As they round Cape Farewell the Irminger Water subducts under the Polar Water.

Adjacent to the coast, the surface water is modified by runoff from Greenland whereby the stratification in the water column increases. During summer, this forms a freshwater baroclinic jet, named the East Greenland Coastal Current, which is only a few tens of kilometres wide but with maximum velocities exceeding 1 m/s (Bacon et al., 2002; Wilkinson and Bacon, 2005). The runoff originates from melting of the Greenland Ice Sheet as well as from precipitation. Kiilsholm et al. (2003) showed that by far the greatest value (i.e., precipitation minus evaporation) over Greenland is found in southeastern Greenland, which could be the main driver behind the East Greenland Coastal Current. If so, the presence of significant, baroclinic coastal currents is limited to southwestern Greenland waters, but this is still an open question, especially due to the lack of observations in northeastern Greenland waters.

The classic time series of temperature and salinity on top of Fylla Bank describes the observed hydrographic conditions off West Greenland. For the temperature curve, the most striking feature is the three cold periods centered around 1970, the early 1980s and the 1990s; all three related to atmospheric forcing (Buch et al., 2004). The cooling in the late 1960s was associated with a large pulse of freshwater that left the Arctic Ocean through Fram Strait and moved rapidly southward in the East Greenland Current and later northward in the West Greenland Current, and it was easily recognized in the Fylla Bank salinity. This event has been labeled the 'Great Salinity Anomaly' (GSA) (Dickson et al., 1988; Belkin et al., 1998), and Curry and Mauritzen (2005) estimated the freshwater flux to be about 0.07 Sv. This value is in agreement with model simulations preformed by Haak et al. (2003) and Olsen and Schmidt (2007).

Interestingly, the freshwater anomaly associated with the GSA is similar to the expected freshwater anomaly due to the forecast maximum melting of the Greenland Ice Sheet. In other words, the natural variability observed today in freshwater fluxes is similar to the expected additional contribution from the Greenland Ice Sheet. However, events similar to the GSA are expected in the future and more importantly, as noted by Vizcaíno et al. (2008), the atmospheric moisture transport is expected to increase by one order of magnitude more than the freshwater increase due to melting of the Greenland Ice Sheet.

In a future climate with increased runoff due to increased precipitation and melting of the Greenland Ice Sheet, the East Greenland Coastal Current would be expected to increase. In a study of climate changes in Greenland waters and surrounding seas, high-resolution regional models were applied to calculate a scenario of the climate for the period 1950 to 2050 (Stendel et al., 2007; Kliem et al., 2007). The regional models were used to downscale a global AOGCM.

The ocean model was forced by the GCM simulation at the lateral boundaries and by a regional climate atmospheric simulation at the surface. Runoff was not included due to melting of the ice sheet directly in the simulations. However, an indirect effect was included by restoring the surface salinity toward the surface salinity of the GCM. Still, Vizcaíno et al. (2008) found that the increase in freshwater flux from the Greenland Ice Sheet is one order of magnitude smaller than the increase in the atmospheric moisture transport under anthropogenic greenhouse gas forcing.

The slight increase in temperature is due to a general atmospheric warming and not caused by the retreat of sea ice, as it is rarely present at Fylla Bank. Moreover, the simulations show a clear freshening at Fylla Bank that is most likely to be a result of the increased hydrological cycle. Note that the major salinity fluctuations are due to major GSA-like events. These natural variations are an order of magnitude higher than the general freshening.

8.4.1.4.1. Fjords

Deep fjords are characteristic elements of the Greenland coastline. They constitute a key element of the land-ocean interface as they connect the ice sheet to shelf waters around Greenland. In addition to tide, runoff and local winds are the main driving forces acting on the upper water masses in a fjord system (Svendsen et al., 2002). Frequently, there is a large climate gradient from the inner parts of the fjords to the outer parts. Also, meltwater from the Greenland Ice Sheet or rivers that drain into the inner parts of the fjords also causes a strong salinity gradient along the fjord axis. Typically, meltwater discharge takes place over a few months when temperatures exceed 0 °C and the discharge often occurs in pulses (Hasholt et al., 2008).

Where a glacier meets ocean water, mass is not just lost by the visually spectacular calving mechanisms. At ice shelves and floating tongues it is known from mass conservation considerations that sub-ice melting can occur at rates exceeding 10 m/y (Rignot and Jacobs, 2002; Rignot and Steffen, 2008). Also, Motyka et al. (2003) proposed a model of freshwater-driven fjord convection that brings in warm saline bottom water and transports cold fresh surface water out of the fjord. Recent observations of high water temperatures (3 to 10 °C) in the immediate vicinity of tidewater glaciers (Motyka and Truffer, 2007; Ritchie et al., 2008; Rysgaard et al., 2008) indicate that the ocean water around Greenland has a large heat content and a corresponding melting potential. Changes in the ocean have been observed to coincide with glacier changes in this system and elsewhere in Greenland (e.g., Howat et al., 2008; Holland et al., 2008).

Numerical simulations of the hydrographic conditions under a warmer climate scenario exist from a northeastern Greenland fjord (Young Sund, 74° N) that is connected to the Greenland Ice Sheet via river runoff. However, no large floating glaciers are present in the fjord system. In a warmer climate, the runoff from land would increase and this would locally change the circulation and influence the biological production (Rysgaard et al., 2003; Bendtsen et al., 2007). The sensitivity of Young Sund to changing runoff shows that in the case of no runoff, the mixed layer thickness is about 9 to 11 m during the summer period due to a weak pycnocline in the upper part of the water column (Figure 8.34).

At a runoff corresponding to only 50% of the normal runoff, the halocline is quickly established, and the mixed layer is about 0.7 to 1.6 m deeper in July and August than under present-day conditions where the mixed layer depth is about 4.0 to 5.2 m in July and August. At a runoff two times greater than in the reference case, the mixed layer becomes 0.3 to 1.1 m shallower during the summer season than in the reference case, and in the extreme case when runoff is 8 times greater, mixed layer depth decreases to about 2.2 to 3.7 m during the summer season.

These sensitivity studies show the importance of the runoff for controlling the depth of the mixed layer. Without a runoff, the mixed layer is controlled solely by wind-induced mixing and buoyancy fluxes at the surface. Even at a moderate runoff of only 50% of the present level, a freshwater controlled mixed layer is established quite early by the end of June, and is only slightly deeper than in the reference case. At the other extreme, when runoff is large, the mixed layer depth is controlled by the strength of the surface forcing on the system, that is, wind and air temperature. Thus, the case with no runoff puts an upper limit on the mixed layer depth of about 11 m in the fjord during the summer season, and, correspondingly when runoff is 8 times the current level this puts a lower limit on the mixed layer depth of about 2 m.

In a future warmer climate scenario, the runoff period might increase significantly, and this would prolong the period in which surface conditions are controlled by freshwater discharge. However, this

would not be expected to decrease the mixed layer depth significantly because the balance between runoff and atmospheric forcing is established within a few weeks.

Whereas mixed layer depth would change little due to increased freshwater runoff, it may alter significantly the transport of saltwater into the fjord from the Greenland Sea due to increased estuarine circulation. Hence, in a future climate scenario primary production is expected to increase due to a combination of increased nutrient import to the fjord and improved light availability as a result of reduced sea ice (Rysgaard et al., 1999, 2003; Bendtsen et al., 2007).

8.4.1.5. Summary

Melting of the Greenland Ice Sheet will affect the physical marine environment on different scales, ranging from local scales to global scales. The most visible effect, which is also the one that attracts most attention in public debate, is the rising water level of the world ocean. The latest assessment from the IPCC (2007) predicts the water level to rise between 0.18 and 0.59 m by 2100. This forecast has been questioned by several scientists, however, because recent data show that global warming seems to accelerate the melting of the Greenland Ice Sheet, which will lead to an increased rise in sea level. Estimates are still rather uncertain but recent studies project a sea level rise of 0.5 to 1.6 m by 2100.

More locally, the most visible effect of the melting ice sheet on the marine environment may be the changes in salinity conditions especially in the Greenland fjords and coastal waters. These may result in stronger gradients between the relatively low-saline surface water and the more saline water below; stronger gradients will have a great impact on primary production.

Although the increased runoff of meltwater from Greenland will reach the areas of deep water formation in the North Atlantic (i.e., the Greenland Sea and Labrador Sea), the freshwater volume from the Greenland Ice Sheet is around one order of magnitude less than that from other freshwater sources (i.e., the Arctic Ocean), and so it seems to be of minor importance to changes in the global thermohaline circulation although this has not yet been fully proven.

8.4.2. The marine environment and marine ecosystems

Almost no research has focused on the effects of increased melting of the Greenland Ice Sheet on marine ecology. Present knowledge on the impact on marine ecology due to climate change in the Arctic region is mainly related to the melting or reduction of sea ice although increased meltwater production will also influence the marine food web especially in the fjords.

8.4.2.1. Primary and secondary production

Overall, predictions of how the marine food web will respond to climate change are challenging due to the non-linear nature of ecosystems. Changes in the phytoplankton influence the copepods that, in turn, influence the phytoplankton through changes in grazing pressure. The same is true for the relationship between copepods and their predators, as well as for relationships further up the food chain. Thus, more research is needed before the essential details of the effects of climate change on marine organisms at the base of the food web can be understood and reliable predictions of ecosystem changes can be made.

In summary, climate-mediated changes of the Greenland Ice Sheet and subsequent changes in physical properties of the marine environment may have large consequences for the succession, composition and production at the base of Arctic marine food web. These changes will propagate up the food web and ultimately affect populations of fish, birds and marine mammals with major consequences for the structure, productivity and exploitation potential of the Greenland marine ecosystem.

8.4.2.2. Important economic species

Greenland waters are important fishing grounds, especially for the Northern shrimp (*Pandalus borealis*) and Greenland halibut (*Reinhardtius hippoglossoides*) fisheries that are essential to the

economy of Greenland (Buch et al., 2004). Northern shrimp, currently the most important marine resource in Greenland, accounts for more than 70% of the total fisheries revenue. Catches have gradually increased to around 150 000 t/y (Kingsley, 2007), with 90–95% harvested off West Greenland. Northern shrimp in West and East Greenland waters are considered separate stocks.

Greenland fish stock composition and stock size is believed to change as an effect of climate change but there are, however, pending research projects to further study and quantify such changes. The effect of increased melting of the Greenland Ice Sheet, however, is believed to be of minor importance compared to the effects of increasing temperature especially on the offshore fishing grounds.

8.4.2.3. Marine apex predators in Greenland

Currently, there are only a few predictions of the direct effects of the melting of the Greenland Ice Sheet on marine apex predators. However, this large-scale event will have cascading effects on the primary productivity and associated species in the marine ecosystem; thus, many indirect effects can be expected.

8.4.2.4. Summary

Generally, climate change has immediate effects on the physical environment of the ocean and in Greenland waters these effects are increased by a strong inflow of freshwater from the melting Greenland Ice Sheet. These changes may have large consequences for succession, composition and production at the base of the Arctic marine food web. The changes will propagate to levels higher up the food web and will ultimately affect populations of fish, birds and marine mammals with major consequences for the structure, productivity and exploitation potential of the Greenland marine ecosystem.

To date, there has been almost no research on the effect of the melting Greenland Ice Sheet on marine ecology; therefore much of the present knowledge on the effects of climate change is related to the melting and reduction of sea ice. Research is therefore needed to quantify how the higher trophic levels of the food chain will be affected by changes in the physical marine environment around Greenland as a result of the increased melting of the Greenland Ice Sheet. To be able to make proper and scientifically based plans and decisions for the future, such research is extremely important to the Greenland society since its economy is highly dependent on the surrounding ocean.

8.5. Socio-economic and cultural aspects of changes in the Greenland Ice Sheet

- Impacts of climate change cannot be studied in isolation from other environmental and social and cultural changes.

- Until now, the observed effects of changes in the Greenland Ice Sheet due to global warming have caused only minor impacts on society. The most marked change observed by the Greenland population has been the retreat of Jakobshavn Isbræ. This has caused an increase in tourism and has increased international attention on Greenland as an icon of global climate change.

- Planned hydroelectric power plants rely solely on runoff from the Greenland Ice Sheet and retreat of the ice sheet, and to be reliable and economical [rentable / viable?] need to be based on reliable projections of the future evolution of the ice sheet.

- Retreat of the Greenland Ice Sheet will open areas for mining but will also result in longer walking distance in relation to hunting.

This chapter evaluates the possibility for projecting socio-economic and cultural impacts on Greenland society caused directly or indirectly by changes in the Greenland Ice Sheet. The chapter describes the modern Greenland society from a historical and contemporary perspective and introduces a number of specific cases that illustrate the propensity for change in a society that is derived from the Inuit

culture; a culture that has survived over a millennium of climate change and external cultural influences.

It is essential to emphasize that impacts of climate change cannot be studied in isolation from other environmental and social and cultural changes. History shows that Arctic societies, like Greenland, have always changed and have always demonstrated a great ability to adapt to change. The implications of climate change have been the strengthening of ongoing socio-economic changes, and have contributed to outlining possible development paths. On the other hand, socio-economic and cultural changes have been reflecting a more or less pragmatic approach both to changes in the environment (including climate change) and the political reality.

8.5.1. History of change and contemporary socio-economic and cultural trends

The shift from an economy based on marine mammals to one based on fisheries has had profound consequences on society, first of all through the sedentarization process. The shift from a cod- to a shrimp-based economy furthermore consolidated an already ongoing process of concentration and centralization, but did not totally restructure the organization of the communities. The present-day diversification both of renewable resource exploitation and the economy in general may contribute to a partial reversal of the former concentration policy by making some of the settlements located near new resources and new economic opportunities (e.g., tourism) a potentially more active part of the economy (Rasmussen and Hamilton, 2001; Hamilton et al., 2003).

Although Greenland has considerable raw material deposits, commercial exploitation has been limited by adverse natural conditions and difficult logistics. Cryolite, coal, copper, marble, zinc, lead and silver as well as gold, olivine, and rubies have all been mined – and the latter three recently extracted. International interest in exploration for minerals and hydrocarbons is increasing and some deposits may prove of economic interest in the future not least because of increasing accessibility due to changes in the Greenland Ice Sheet.

Hydroelectric power is a new potential resource, not for energy export but for domestic consumption and for energy intensive activities such as aluminum production. It is important to note that hydroelectric power is based entirely on water basins supplied by meltwater from the Greenland Ice Sheet.

8.5.2. Implications of changes

The new expectations for large-scale mineral and energy resource production as well as the dependence on fisheries and hunting activities are influenced by changes in climate and associated changes in the Greenland Ice Sheet and sea-ice conditions. Renewable resources are affected through their interactions with the environment, while non-renewable resources are affected by changes in accessibility and transportation.

The smaller settlements will experience the immediate effects of these changes because renewable resources are still of significance to them and because access to resources and to settlements will be altered by changes in accessibility, abundance and geographic distribution of these resources, as well as by changes in their economic importance (Poppel, 2007). Tourism has become an alternative source of income for many local communities, enabling a positive interaction between new economic opportunities and traditional activities that are less affected by change than exploitation.

At a longer perspective, Greenland's economy is shifting toward increased dependency on non-renewable resources and in so doing will be highly affected by the ongoing changes, including those affecting access to resources and transportation. The decisive force will be the population's ability to adapt to the changes and to continue to improve the level of education (Hamilton and Seyfrit, 1993; Paldam, 1994; Rasmussen, 1997, 1998, 2002; Winther, 1999, 2001; Zelarney and Ciarlo, 2000; Duhaime et al., 2001; Hansen, 2003; Nellemann and Vistnes, 2003).

8.5.3. Influences

8.5.3.1. Accessibility and impact on communities

Changing ice conditions, both on land and at sea, are important factors influencing everyday life in most communities in Greenland, mostly by determining the accessibility of the water bodies. Access to the sea, and sea transport to and within Greenland, are determinant factors for hunting, fishing, exchange of goods and tourism, which is becoming increasingly important for many communities.

Hunting terrestrial mammals is contributing both to subsistence and informal economic activities of many Greenlanders (Dahl, 2000; Poppel and Kruse, 2009). Muskox (*Ovibos moschatus*) and caribou / wild reindeer (*Rangifer tarandus*) are hunted by licensed commercial hunters and by sports hunters in the autumn, when as much as a fifth of the population goes hunting. Changes in the position of the Greenland Ice Sheet margin may affect hunting activities by changing the routes of rivers draining meltwater from the ice, and in conjunction with rising temperatures may affect the migration patterns of the game. In cold weather, the animals move toward the coast, where they become more accessible to hunters, while in warmer weather, the animals move toward cooler conditions closer to the ice margin in order to reduce insect harassment. As a result, the walking distance of the hunters to the animals becomes increasingly difficult to cope with, first while searching for the animals, and then especially when the meat is carried on the shoulders back to the boats anchored along the coast.

8.5.3.2. Possible consequences for new activities

Reduction of glaciers, ice sheet and sea ice in and around Greenland will increase accessibility to new areas with the potential for commercial exploitation of minerals and energy resources. A significant side effect to the retreat of the ice sheet margin will be the increased volume of meltwater that can be harnessed for producing hydroelectricity. This potential is already included in the new economic strategies for Greenland. Over the past 50 years, the hydroelectric power perspectives in Greenland have been mapped and evaluated as a potential contribution to the development of the national economy. The hydroelectric power exploitation has already emerged in the form of three hydropower plants: A 30 MW plant near Nuuk, a 1.2 MW plant in Tasiilaq, East Greenland, and a 7.2 MW plant near Qaqortoq, South Greenland. A 15 MW hydropower plant near Sisimiut, West Greenland, is under construction and will become operational in 2010.

Several other hydropower plants are in the planning process, now aiming at providing power to new economic activities connected to energy intensive large-scale raw material processing. The first project is the supply of energy to an aluminium production plant currently planned at Maniitsoq, West Greenland. Even though Greenland does not have the ore resources needed for aluminum production, import of ore (i.e., bauxite) from Australia or South America to Greenland could be a viable economic activity given the access to large and continuing amounts of meltwater from the Greenland Ice Sheet and the derived hydroelectricity.

A key question, however, is the stability of the water supply to the new power plants as the ice sheet margin retreats due to melting. Increased amounts of silt and sand in the meltwater may fill dammed lakes and affect the viability of the dams, eventually requiring additional reconstruction in order to maintain the water flow needed. Or worse, the bedrock topography exposed by the retreating ice may cause new meltwater drainage patterns, leading to a partial or complete drying out of the water supply to the hydropower plants (Anon, 1994; Udvalget om socioøkonomiske virkninger af olie- og gasudvinding samt mineralindustri, 1997; Rasmussen, 1997, 2005; Keddeman, 1998; Storey and Hamilton, 2003).

8.5.4. The Ilulissat case

The Jakobshavn Isbræ (i.e., the Ilulissat Glacier) at the head of the Ilulissat ice fjord is one of the world's fastest moving glaciers producing more than 10% of the total discharge of icebergs from the ice sheet (Figure 8.35). During the culmination of the last Ice Age 21 000 years ago, an ice sheet that reached out onto the margin of the continental shelf covered the present-day ice-free land bordering the Greenland Ice Sheet.

After the end of the Ice Age, the melt-back of the ice started about 11 700 years ago and, during a warm climatic period 8000 years ago, the ice rapidly regressed causing the glacier front to retreat inland to a position more than 50 km east of the present grounding line. With the onset of the Little Ice Age, the glacier front again advanced, and in 1851, a maximum extension was recorded in the middle of the ice fjord close to the former Inuit settlement Qajaa. Following this advance, the ice front stabilized and a slow regression prevailed until the front suddenly collapsed in 2002/2003 causing the floating ice tongue to withdraw more than 15 km eastward in just a few years. This rapid withdrawal ceased in 2007, when the ice front reached the grounding line.

8.5.4.1. Inuit settlements, cultural changes and role of the Ilulissat ice fjord

The glacier and the ice fjord have had an immense impact on the population of the region during the history of human occupation. The glacier produces a large amount of nutrient-rich meltwater that facilitates the primary production in the fjord. The constant input of nutrients thus forms the basis for a rich fauna in the fjord, which has provided excellent hunting and fishing conditions for people inhabiting Disko Bay through the ages. Prehistoric hunting areas and settlements were established and abandoned as the location of the glacier front changed. In this way, the Ilulissat area illustrates the interplay between an ever-changing glacier environment and the development of human occupation.

Many archaeological findings testify to the long history of settlement in the Ilulissat ice fjord region, where hunting and fishing have always been the basis for life. The Saqqaq people reached West Greenland from Alaska 4500 years ago and settled in the Ilulissat area where they lived for 1500 years before they disappeared. West Greenland was then uninhabited until a new wave of paleo-eskimo settlers, the Dorset people, arrived 2800 years ago. They occupied the region for 800 years before they disappeared leaving the Ilulissat region and Greenland uninhabited for the next 1000 years. By 1200 AD, people of the Thule culture settled in the region and established two flourishing communities, Sermermiut and Qajaa, along the ice fjord. These settlements are among the largest prehistoric Inuit settlements discovered in Greenland and, together with the dense pattern of smaller settlements in the area, indicate that the living resources around the ice fjord were rich, stable and plentiful over long periods of time (Grønnow and Meldgaard, 1988).

8.5.4.2. The Ilulissat ice fjord and present-day socio-economic effects

In 1741, the Danish colony of Jakobshavn (today Ilulissat) was established and this led to the depopulation of the rich Sermermiut settlement around 1850. Today, Ilulissat is the third-largest town in Greenland with a population of more than 4000. Half of Greenland's coastal fisheries are now centered here and the town is also home to one of the country's largest fish-processing factories. Greenland halibut are plentiful throughout the fjord system and are subject to fishing based on a quota system. Long-line fishing for Greenland halibut from the sea ice in the Ice Fjord has been an important part of the fishing tradition through centuries and it is still taking place.

The abundance of resources and the adaptability of the traditional culture to the changing conditions of the glacier and winter sea ice have been crucial to the prosperity of the social and economic life of the region. Although the remarkable present changes in the ice will constitute new challenges to economic and social activities in the region, the conditions may now be different, less predictable and more challenging than before: Icebergs are smaller and the immense discharge by the glacier in combination with a rise in sea temperature, has caused a considerable reduction in sea-ice cover during winter.

This may imply profound changes for the production of nutrients – the basis for biological life in the region – and may lead to smaller stocks of Greenland halibut and northern shrimp. These species are fundamental, not only to human harvesting, but also to the seals and whales that form a significant part of the daily food supply for humans and dogs. Also, the distribution of seabird species may be affected. Other species may enter the region offering new and potentially prosperous options for economic activities. Such challenges will be met as before.

The adaptations of today are, however, far more complicated. Modern and economically viable harvest of renewable resources is highly industrialized, and to establish new production will be costly, requiring national initiatives as well as international involvement and regulations applied to managing, accessibility and rights. The local capacity for adaptation will be met by implications of today's global interaction. Changes in sea-ice cover, especially a reduction in winter sea-ice extent and thickness will inevitably alter marine transport conditions. Reduced sea ice may offer new transportation routes and prolonged navigation seasons, but will also increase marine access to natural resources and, accordingly, the potential for disturbance and for more efficient fisheries. Although the changes in sea ice may lead to significant changes in the socio-economic life of the region, it is still premature to state the nature of these socio-economic changes. The unpredictability of climate change and its local implications make planning, decision-making and management even more challenging. In spite of its richness, Arctic nature is vulnerable and poor management may lead to serious and irretrievable damage.

8.5.4.3. The Ilulissat ice fjord and its impact on science, politics and tourism

The Ilulissat ice fjord and its glacier is one of the world's best-studied glaciers with the longest observational record. The ice fjord has attracted many scientists through the centuries, the first being H.J. Rink who studied and described the glacier and the Greenland Ice Sheet in the mid-19th century. The studies were instrumental in formulating the important theory of the Quaternary glaciations. Today, the ice fjord is still the subject of detailed international research because of its exceptional manifestation of changes.

The Ilulissat ice fjord was nominated a UNESCO World Heritage site in 2004. The World Heritage List uses the name 'Ilulissat Icefjord' for the entire world heritage site, which includes land areas, the glacier front and parts of the Greenland Ice Sheet adjacent to the ice fjord, recognizing the uniqueness of the area and the extensive interplay between nature and culture.

By entering this prestigious list, tourism has virtually exploded in the ice fjord region. This has made Ilulissat Greenland's leading tourist town and the area now hosts more than 50% of the total number of tourists visiting Greenland. Thus, the ice fjord is by far the main attraction and cruise ships arrive in increasing numbers during the summer to let their passengers visit the fjord.

It is not just tourists that are congregating in the area. The Ilulissat ice fjord has also attracted a large number of politicians from around the world who wish to experience firsthand the global climate change as witnessed here. The ice fjord has now entered political discussions in a way that has never been seen for any part of Greenland before (Mikkelsen and Ingerslev, 2002).

8.5.5. Conclusions

Both local and global effects of the ongoing changes in Earth's climate will inevitably affect Greenland and its population. This chapter has attempted to assess the extent to which changes in the Greenland Ice Sheet will be a major factor in shaping the future society of Greenland. The response of Greenland society to past changes in climate has also been reported and has compared these changes to the effects of societal dynamics. The history of the Thule culture, which gradually transformed into the present-day society in Greenland, has been one of constant adaptation to changes in climate and increased interaction with European culture.

Until now, the observed effects of changes in the Greenland Ice Sheet due to global warming have caused only minor impacts on society. The most marked change observed by the Greenland population has been the retreat of Jakobshavn Isbræ. This has caused an increase in tourism and has increased international attention on Greenland as an icon of global climate change.

In the coming years, changes in drainage patterns around the Greenland Ice Sheet may affect coastal waters and thereby cause changes in the distribution of marine fish stocks and game species. Future changes in drainage patterns may also be a major concern for large infrastructure investments with decadal to centennial life spans, such as hydroelectric power plants and transmission lines, roads,

airports and seaports. This chapter has not considered effects from changes in permafrost, although these may have considerable economic implications; the issue of permafrost is considered as separate from that of the ice sheet. Based on the historic observations and the strong drive for change in Greenland society today, it is likely that the consequences of changes in the Greenland Ice Sheet will be minor compared to socio-economic, cultural and demographic influences generated within the Greenlandic and international societies.

8.6. Major knowledge gaps and recommendations

8.6.1. Introduction

New research results from the Greenland Ice Sheet often lead to the identification of new gaps in knowledge, although they may become more specific over time (see Table 8.6). For example, research projects initiated under the International Polar Year 2007/2008 should eventually lead to the identification of new gaps in knowledge or to changes in the existing gaps in knowledge.

Table 8.6. Major gaps in knowledge and recommendations.

Major gaps in knowledge	Recommendations
Large variability between presently used SMB models	Targeted in situ SMB data with which to calibrate/validate models
Incomplete understanding of SMB processes	Systematic studies focused on improvement in process understanding
Large spatial variability in SMB at the ice sheet margin	Better sub-grid scale quantification of the key processes; refreezing, albedo and blowing snow
Large error in determination of surface annual accumulation	Improved determination of snow densification
Discrepancy in estimating the mass change from satellite gravity measurements	Establish a community wide recognition on which processing method provides the most reliable method for tracing mass changes
Incomplete estimate of the mass loss from marine terminating outlets	More systematic mapping of ice thicknesses from outlet glaciers
Incomplete determination of the annual cycle in SMB from satellites	Sampling sufficiently frequent to capture the annual cycle in accumulation, melt and iceflow
Lack of ice discharge determination from ice streams and narrow outlets	Development of high resolution sampling methods from satellites to capture basin scale changes
Large seasonal as well as interannual variability in marginal ice fluxes	Improved theoretical understanding of the basal hydrology and its impact on ice flow
Basic understanding of coupling between ocean and marine based fronts	Theoretical understanding of the calving and basal melting at marine margins where changes occur
Lack of thickness determination around highly dynamic outlets	Develop methods to retrieve ice thicknesses at heavily crevassed margins
Incomplete understanding of flow properties deepest in the ice sheet	Determination and validation of the thickness of the basal temperate layer as well as deep ice temperatures
Incomplete cryospheric representation and determination in climate models	Validate climate models more systematic with observations of surface climate and mass balance of the ice sheet
Lack of full coupling between surface dynamics and the atmosphere	More accurate incorporation of surface albedo and snow microphysics in climate models
Major dynamical features like ice streams lack representation in ice sheet scale models	Better model resolution as well as process understanding leading to well parameterized models
Several-fold changes in velocity variations at marine based outlet glaciers	More complete determination of process leading to improved parameterizations of basal and marginal (ocean) linked process

8.6.1.1. Assessment of the impacts

Earlier assessments identified some of the major gaps that are currently being addressed. The Arctic Climate Impact Assessment (ACIA, 2005) focused on improving projections of future changes and identified several major gaps in knowledge that are also valid for the Greenland Ice Sheet:

- a need to improve understanding of albedo changes and feedback mechanisms
- the need for studies of outlet glacier dynamics with an emphasis on their potential for triggering persistent and rapid thinning
- a need for improving ice dynamic models for determining the long-term response of the ice sheet to past climate change
- a need for improving parameterization and verification of internal accumulation models
- a need to improve understanding of the relationships between climate change, meltwater penetration to the bed, and changes in iceberg production.

Uncertainty in scenarios concerning how glaciers, ice caps and ice sheets contribute to sea level rise over decadal to centennial time scales needs to be reduced. This requires an assessment of the Greenland Ice Sheet and its stability and vulnerability to climate change, including sudden and potentially irreversible climate change. The Climate and Cryosphere Project (CliC), established in March 2000 to stimulate, support, and coordinate research into the processes by which the cryosphere interacts with the rest of the climate system, recommends that an insight is needed into the highly non-linear links between the various components of the cryosphere and their likelihood for producing significant global change. CliC recommends that the long-term ice sheet monitoring system is improved in order to make realistic representations (modeled as well as observational) of spatial and temporal variability in the surface mass budget in areas sensitive to sea level change (Climate and Cryosphere Project, 2007).

The IPCC's Fourth Assessment Report (IPCC, 2007) stated that realistic scenarios of the Greenland Ice Sheet SMB require a resolution exceeding that of the current AOGCMs used for long-term climate experiments. Present climate models do not generally include a representation of the refreezing of surface meltwater. All models predict an increase in snow accumulation, but there is much uncertainty in its size. In projections for Greenland, an increase in ablation is important but uncertain, being particularly sensitive to temperature change around the margins. The main uncertainty is the degree to which outlet glaciers having a marine termini respond to climate change. Further accelerations in ice flow, of the kind recently observed in some Greenland outlet glaciers (and in West Antarctic ice streams), could increase the ice sheet loss of mass substantially, but quantitative projections cannot be made with confidence.

8.6.1.2. Summary of earlier assessments

The research needs presented in the ACIA, CliC and IPCC assessments of the Greenland Ice Sheet vary in their degree of detail, but all focus on the importance of making more precise scenarios and filling gaps in knowledge. There is a common agreement, that improved understanding of processes is needed in order to determine meltwater runoff more precisely. Also, the rapid increase in ice discharge from the outlet glaciers observed over the past decade calls for a better understanding of processes as well as a better formulation of theory before models quantifying the observed increase in ice flux can be developed. The three assessments all call for better monitoring capabilities and a more precise determination of present state of balance in the Greenland Ice Sheet before more reliable scenarios can be made.

8.6.2. Gaps in knowledge of surface mass balance

Significant differences exist between different estimates of the components of the SMB derived from numerical modeling and downscaling of climate re-analysis data. For example, there are important differences in distribution and total accumulation in the 'wet' southeastern Greenland. The differences between model estimates can be as large as the net SMB for a given year and the standard deviation of the differences is of the order of 100 Gt. Although values for total runoff may agree between models,

the individual terms included in the calculations of total runoff vary much more widely, particularly when comparing melt and refreezing. Total differences are of the same order of magnitude as the inferred increase in mass loss due to changes in ice dynamics over the past decade (Rignot and Kanagaratnam, 2006). The differences suggest that mass budget calculations are seriously hindered by uncertainties in SMB.

Direct observations of increasing surface elevation in the regions above 2000 m are not confirmed by direct observed accumulation changes due to very high variability in the records. RCM reconstructions of precipitation do not predict increased values.

Uncertainties in SMB originate from a range of sources related both to lack of in situ data and incomplete process understanding. To reduce uncertainties to a level useful for mass budget calculations and reliable predictions will require concerted effort. Especially, (i) to collect more targeted in situ data to calibrate and validate models, (ii) to improve process understanding, and (iii) to model key processes such as refreezing, surface albedo, blowing snow and sub-grid scale effects. Sub-grid scale effects are particularly significant in coastal regions due to higher surface relief, which makes ablation observation uncertainties difficult to incorporate into SMB model validation. The most important parameters for controlling runoff in SMB modeling, however, are surface albedo and meltwater refreezing.

8.6.3. Gaps in knowledge of net balance or components thereof

At present, the best estimate of the mass of the Greenland Ice Sheet is probably provided by GRACE (see Section 8.2.4.2.3). However, different processing methods result in different results. One method of averaging the total mass of the Greenland Ice Sheet over 30-day periods and calculating seasonal loss (Velicogna and Wahr, 2006) provides a mass loss of 211 Gt/y, while another uses different processing algorithms for analyzing the distance between the two satellites (Luthcke et al., 2006); this second approach provides a divergent mass loss of 154 Gt for 2007 (Witze, 2008). Such discrepancies reflect ongoing discussion on which method is most valid for detecting mass changes in the Greenland Ice Sheet using GRACE and the newer publications seem to have a better agreement of results.

Measuring volume change by altimetric methods such as laser (IceSat) or by radar (CryoSat) provides another tool for detecting changes in mass balance. However, connecting volume change to mass changes requires insight into densification processes of the surface snow and firn, which also has a strong seasonal dependence when the surface temperature reaches the melting point. Moreover, it remains difficult to produce a measuring accuracy any better than a few tens of centimetres, and this also renders it impossible to capture the seasonality of the parameters.

Satellite radar techniques measuring ice displacements and ice fluxes near calving glacier fronts are unique in providing a means for detecting iceberg production when the ice thickness is known. However, the lack of measured ice thicknesses from all outlet glaciers and the difficulty of measuring ice movement in the interior of the Greenland Ice Sheet constrain this technique.

The various techniques used to estimate the state of balance of the Greenland Ice Sheet differ in their methodology for detecting changes in specific components of the system; which means that they do not return results that are mutually compatible. Closing the gaps between the results for different techniques will make an important contribution toward a more consistent understanding of change in the ice sheet. Filling these gaps in knowledge will require common or compatible temporal and spatial sampling as well as an understanding of the annual cycles in mass balance, including its most important parameters: Accumulation, melt, runoff, and ice flow. Temporal sampling will need to be frequent enough to capture the annual cycle, or designed to mitigate the effect of interannual variations in this cycle on annual estimates.

Spatial sampling capturing variation at a drainage-basin scale would allow for a more direct comparison of changes in discharge and measures of ice sheet mass or volume. Improved sampling of surface height with new satellite missions may provide altimetric determinations that more faithfully reflect the impact of annual cycles on surface height. However, in order to secure significant progress

in understanding densification processes, observation and modeling of the surface snow are also required.

8.6.4. Gaps in knowledge of ice dynamics

A better understanding of basal boundary conditions (i.e., sliding laws) is needed. This involves a more precise understanding of the mechanisms of basal motion in order to derive good parameterization, which accounts for seasonal and longer-term variability. An improved understanding of basal boundary conditions is inherently linked to an understanding of subglacial hydrology, which is assumed to drive the present speed-up of calving outlets. For instance, subglacial hydrology and the complex connection between surface hydrology (supraglacial lake drainage) and the bed are integral to basal flow, which can dominate ice flux near the margin. Modeling the ice sheet hydrology, however, is still in an early stage of development and has not yet been coupled with ice dynamics.

Concerning the ice-ocean interface, gaps in knowledge are mainly related to the assessment of melting rates under floating glacier tongues, where the ocean and heat circulation in combination with freshwater from melting ice drive the circulation and impact on the calving front (Motyka et al., 2003). Generally, iceberg calving and basal melting at marine margins are difficult to measure and to simulate, and these processes are absent or overly simplified in current ice sheet models.

Observational gaps mainly relate to determining ice thickness at heavily crevassed outlet glaciers, due to difficulties retrieving radar signals in these regions. Also, ice temperatures and, in particular, thickness of the basal temperate layer are not well known, which in combination with ice fabric (Lüthi et al., 2002) show a strong dependence on ice deformation rates and hence dynamics of the outlets.

8.6.5. Gaps in knowledge of ice sheet and climate modeling

Models of meteorological processes, ice sheet mass balance, and ice sheet dynamics are constantly improving, but there are still many uncertainties associated with simulating the Greenland Ice Sheet and its likely response to climate change.

8.6.5.1. Climate models

Many climate models do not resolve the steep topography of the Greenland Ice Sheet margins very well, limiting model skill in simulating orographic precipitation and ice sheet ablation in this zone. High-resolution meteorological models such as PMM5 (e.g., Box et al., 2006) overcome this problem, but there are still outstanding issues with regard to temperature biases, simulation of mass balance processes such as sublimation and boundary layer winds, and the large-scale boundary conditions for regional climate modeling in future climate-change scenarios.

The dominant precipitation and ablation processes can also be event-driven in Greenland (i.e., resulting from a small number of extreme synoptic weather events), and these processes are well captured in simulations that are forced by historical (i.e., re-analyzed) climatology but do not necessarily arise in climate model simulations strong in predictions of the mean state; and they do not always represent the full range of weather variability.

Probably the most important boundary conditions for climate model simulations of the Greenland Ice Sheet are the surface properties (i.e., the physical representation of snow in the models). The inherent instability of snow crystals on the ground, the ever changing meteorological conditions, and the feedback between snow and atmosphere via snow albedo makes the snow-atmosphere coupling an important and highly dynamic one. New progress in snow physics theory and parameterizations needs to be incorporated into climate models.

8.6.5.2. Glaciological models

Ice sheet models now operate at resolutions of 5 to 20 km for the Greenland Ice Sheet, representing the overall ice sheet geometry quite well but still not capturing details of ice margin positions or some of the major fjords and outlet glaciers. This means that it is still difficult to compare model simulations with geological reconstructions of recent (i.e., 20th century) ice sheet changes. In addition, it is difficult to compare modeled ice discharge with observations, with the more distributed (i.e., less channelized) ice flux in the model. Major ice dynamical features such as the glaciers at Ilulissat and Storstrømmen (also known as the Northeast Greenland Ice Stream) are known to account for a large percentage of ice discharge from the Greenland Ice Sheet, but these features are generally absent in ice sheet models. This is partly an issue of model resolution and partly an issue of complex dynamical processes that are presently too poorly understood to be well parameterized in models.

Ice sheet models are presently insensitive to ocean forcing, in contrast with recent observations of significant (several-fold) velocity variations in marine-based outlet glaciers, most likely to have been triggered by ice-ocean interactions and thinning at the grounding line. Models still cannot simulate this interannual flow variability, so it is unclear whether they are able to provide a credible estimate of the response time of ice sheet change to atmospheric or oceanic forcing.

8.6.6. Perspectives

During the 20th century, the mass discharge from the Greenland Ice Sheet was almost equally divided between meltwater runoff and iceberg discharge (Reeh, 1994) – making the two terms equally important. The magnitude of the present-day balance is uncertain due to lack of coherence between the different estimates relying on different methods. Closing the gaps between the results is a prerequisite for reliable estimates of the present balance. Furthermore, a firm knowledge about the present balance is needed in order to generate reliable future scenarios.

Making specific recommendations on research needs to close the gaps in knowledge requires detailed insight into the problems. However, detailed insight is still lacking for many of the major knowledge gaps, such as the processes underlying increased ice discharge and meltwater refreezing. Instead of reducing uncertainty about changes in the mass balance of the Greenland Ice Sheet, the new observational methods from space have revealed to the research community the incomplete understanding of the fundamental dynamics of calving glaciers. This lack of understanding may explain why gaps in earlier assessments are still unresolved in the present-day assessment of the Greenland Ice Sheet. Closing such gaps in knowledge, where there is a lack of fundamental understanding of the processes, demands the development of theory and methodology to validate the theory against field measurements.

Most changes appear in the marginal regions of the Greenland Ice Sheet where data coverage is most scarce due to logistical access constraints. Moreover, thin ice in these regions, combined with a complex terrain, results in highly variable ice dynamics, melt rates, and accumulation patterns. This calls for high spatial resolution modeling in combination with enough in situ measurements to capture adequate detail and to gain sufficient understanding of the present thinning process.

New satellites have triggered a quantum leap in surface observations of the Greenland Ice Sheet. However, in situ measurements are still progressing at a slow pace due to resource-demanding fieldwork. The development of new, and less labor-intensive, measuring techniques could potentially advance the process understanding over larger areas. A combined effort in advancing process theory, field methods and coverage as well as modeling efforts could facilitate major advances in understanding the present balance of the Greenland Ice Sheet and, hence, help create less uncertain future scenarios.

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11.1. Synthesis of Feedbacks and Interactions: From the cryosphere to the climate system – effects over various spatial and temporal scales

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Summary

Changes in Arctic climate are a result of complex interactions between the cryosphere, atmosphere, ocean and biosphere. The present assessment has substantiated and quantified many of the linkages, processes, interactions and feedbacks between the cryosphere and climate across a range of spatial and temporal scales revealing that:

- In the cryosphere-climate system more feedbacks are identified as positive and result in warming than are identified as negative and result in cooling.
- Several positive feedbacks are finite: warming due to reduced albedo will no longer increase in areas that have totally lost snow or ice cover.
- Feedbacks operate at different spatial scales. Many of the feedbacks, such as those operating through albedo and evapotranspiration, will have significant local effects that together could result in a global impact. Some processes such as carbon dioxide (CO_2) emissions are likely to have very small global effects but uncertainty is high. Others, such as subsea methane (CH_4) emissions, could have large global effects.
- Some cryospheric processes in the Arctic have teleconnections with other regions, for example, the loss of sea ice north of Eurasia may result in a cooling effect over eastern Asia, and changes in snow cover affect atmospheric circulation. Conversely, major changes in the cryosphere have been largely a result of large-scale processes, particularly atmospheric and oceanic circulation.
- There are also interactions between the elements of the cryosphere acting through the atmosphere and ocean.
- The cryospheric components of the Arctic play a pivotal role in freshwater generation, its intra-Arctic storage, and routing to the North Atlantic where it can produce an important feedback to regional and global climate. With continued climate warming it is highly likely that the cryospheric components will play an increasingly important role.
- Terrestrial snow cover, sea ice and permafrost are involved in multiple temporal and spatial feedback regimes whereas land ice is involved in fewer feedbacks.
- The overall net effect of all the feedbacks is difficult to assess because of the variability in spatial and temporal scales over which they operate.
- General circulation models (GCMs) do not include all major feedbacks, in part because they do not include all processes that lead to feedbacks (e.g., freshwater runoff from glaciers and ice sheets). Furthermore, the feedbacks may not be accurately parameterized in the models. The lack of full coupling between surface dynamics and the atmosphere is a major gap in current GCMs.

11.1.1. Introduction

Feedback processes are a phenomenon that has been known for more than a century. They are responses to a driving mechanism that subsequently accelerate (positive feedback) or retard (negative feedback) the original driving process. At the end of the 19th century, Arrhenius (Bolin, 2007) described the classic feedback whereby increased air temperature leads to an increase in the amount of water vapor in the atmosphere, which in turn leads to additional atmospheric warming. More recently, Curry et al. (1995) detailed the sea ice-albedo feedback mechanism, including snow cover, melt ponds, and leads. Overpeck et al. (2005) took a broader approach to the analysis of Arctic feedbacks by including all major components of Arctic climate: sea ice, permafrost, terrestrial ice, thermohaline circulation, terrestrial biomass, and marine primary production, as well as population and economic production. Francis et al. (2009) used a similar presentational approach to distil the system into its fundamental physical and human parts, document the key relationships between those parts, and identify the feedback loops. Both Overpeck et al. (2005) and Francis et al. (2009) examined the changes in feedback loops under the scenario of continued warming and a greatly reduced permanent sea-ice cover.

With the vast amount of recently published information on observed and projected trends in the cryosphere, there is a need to re-evaluate the feedbacks between the changing cryosphere and climate. There are many feedbacks from the cryosphere to the climate system: some are direct, but others are complex and indirect; for example, climate effects on the cryosphere that then affect ecosystems that in turn affect the climate system. Some of the complexity and variety of interactions between the atmosphere, ocean, and cryosphere is illustrated in Figure 11.1. Further complexity arises owing to scaling issues: individual feedbacks operate over different time scales and their effects can vary from local to global spatial scales. Furthermore, some feedbacks are negative and result in climate cooling whereas others are positive and lead to warming. Consequently, although some feedback mechanisms have been known for a long time, an overall assessment of the net effect of many different potential feedbacks on climate change has not yet been achieved.

Figure 11.1. Local, regional, and global feedbacks and processes related to changes in the Arctic cryosphere. The numbered yellow circles refer to impacts and processes: (1) Melting and retreating snow cover increases radiation absorption, a radiative feedback. (2) Melting of large ice sheets contributes to sea-level rise and the freshwater flux with potential effects on thermohaline circulation and global climate. (3) Retreating sea ice contributes to increased radiative absorption (ice-albedo feedback) and heat and moisture fluxes to the atmosphere, which impact cloud cover. (4) As permafrost degrades, CH_4 production increases. With wetland drying, CO_2 emissions increase, and the atmosphere warms over time. (5) Thawing permafrost changes geomorphic / geochemical processes and fluxes. (6) Increasing precipitation plus melting snow and ice increases river flow and changes the freshwater flux. (7) Shrinking lake-ice cover has ecological impacts generally leading to greater productivity but negatively impacting surface transport. (8) Changes in the magnitude and timing of snowmelt runoff and river-ice processes have both positive and negative impacts. (9) Retreating glaciers initially increase runoff but lower flows eventually result as ice masses diminish. (10) Changes in cloud cover affect the surface radiation budget, which impacts sea ice, in turn affecting cloud cover (cloud-radiation feedback). Source: after Prowse (2009).

This section summarizes the feedbacks associated with a changing Arctic cryosphere presented elsewhere in this report, describes other feedbacks and interactions that span the various cryospheric components, and provides a preliminary assessment of their relative magnitudes. However, the calculation of the net effects of all feedbacks requires complex modeling that remains a priority for future research.

11.1.2. Feedbacks

11.1.2.1. Greenhouse gases

11.1.2.1.1. Carbon dioxide (land, freshwater, marine)

Carbon dioxide is a greenhouse gas with a radiative forcing of 1.66 W/m^2. The adjustment time of CO_2 in the atmosphere is approximately 100 years (Forster et al., 2007). It is exchanged between the atmosphere and the biosphere through the processes of photosynthetic capture by green plants, short-term autotrophic respiration by plants, and long-term heterotrophic respiration of dead plant material by microbes. A mismatch, particularly in wet areas, between the rate of fixation of atmospheric CO_2 and its release from the biosphere has lead to a net accumulation of carbon in the Arctic for thousands of years with considerable amounts preserved in permafrost (Chapter 5, this volume). Around 44% of the world's near-surface labile soil carbon is found in Arctic soils (McGuire et al., 2009; references in Chapter 5, this volume), which is equivalent to approximately twice as much carbon as is currently present in the global atmosphere (Anisimov et al., 2007; McGuire et al., 2009; Tarnocai et al., 2009). This carbon store is sensitive to climate warming.

The major cryospheric controls on carbon *capture* from the atmosphere are the duration of the snow-free period that determines when photosynthesis can occur efficiently, the timing of the snow-free season (earlier snow thaw allows plants to harvest high levels of solar radiation whereas solar angles are low and radiation is reduced at the end of the snow-free period; Callaghan et al., 2004a) and damage to vegetation from extreme winter warming events that thaw snow briefly in mid-winter

(Bokhorst et al., 2009). Euskirchen et al. (2006) estimated increased carbon drawdown at about 9.5 g C / m^2 / y for each day that the growing season was projected to increase between 2001 and 2100. Recent analyses from visible satellite imagery of changes in snow-cover extent show that decreases are more marked in the spring period than in the period of snow-cover onset, and that in May-June there has been an average decrease of 18% over the 1966 to 2008 period, thereby extending the snow-free season (Chapter 4, this volume).

The major cryospheric controls on carbon *release* from the soil are thawing permafrost (Chapter 5, this volume) and soil warming during extreme events such as fire. Experimental warming of soils (Dorrepaal et al., 2009) and observations of areas that are experiencing different degrees of warming (Lee et al., 2009; Schuur et al., 2009; Vogel et al., 2009) show that the Arctic is already losing some of the carbon that has been stored for thousands of years. Extrapolations from measured carbon emission rates suggest that 4.5 to 6.0 kg/m^2 (or 9.5–13%) of the soil organic matter carbon pool could be lost on a century time scale (Schuur et al., 2009). Currently, widespread permafrost thawing throughout the Arctic has not been reported, although permafrost temperatures are generally rising. However, recent projections of permafrost temperatures show thawing to be widespread throughout the southern regions of the Arctic by 2090 (Chapter 5, this volume).

Overall, models suggest that the Arctic will remain a weak sink of carbon during warming, mainly because of increased growth and extent of more productive vegetation in the future (Callaghan et al., 2004b; Sitch et al., 2007; McGuire et al., 2009). However, both current and future carbon sink activity can be reversed by short-term disturbances such as forest fires and insect pest outbreaks, and long-term changes. The long-term changes include thawing permafrost and altered hydrology related to permafrost dynamics (Chapter 5, this volume) and changes in snow regime (Chapter 4, this volume) that lead to drying and plant water stress. Also, the models exclude disturbance due to human activities such as deforestation at the treeline. Although the future feedback sign and strength status for CO_2 in the Arctic remain uncertain, it is likely that century-long processes of negative feedbacks from increased CO_2 sequestration will be interspersed with episodic releases following disturbances. Some of these disturbances (such as forest fires) will also intensify the positive feedback arising from black carbon deposition on snow and ice through long-range transport on an annual time scale.

The Arctic Ocean could be a significant sink for carbon as sea ice retreats and the ocean warms. Also, the oceans are sinks for carbon over geological time scales as dead organic matter accumulates in deep ocean sediments and carbonate in the shells of some plankton species becomes calcareous rocks. Currently, there can be mismatches between algal blooms at the retreating ice edge (now over the continental shelves) and the zooplankton and bacterio-plankton that feed on them: the algae that are not eaten fall onto the continental shelves and are available for the benthos (creatures living on the seabed) to feed on or form sediments. As the ice edge retreats away from the continental shelves during climate warming, more blooms could occur, which might lead to more organic material falling to the deeper ocean bed off the continental shelves and accumulating there. However, this negative feedback to climate warming might be moderated by ocean acidification. The pH of the ocean is currently decreasing as more CO_2 is drawn down from the atmosphere: the slightly more acidic conditions that result affect the species that use calcium to build cell walls causing less calcium to be deposited onto the ocean bed.

Freshwaters (lakes, ponds, rivers) feed back to the atmosphere through processes that occur in their immediate vicinity (e.g., carbon fluxes, albedo, evaporation) and downstream where the waters enter the Arctic Ocean (see section 11.1.2.3). As temperatures rise, the number of days of ice cover decreases and more heat is absorbed by the open water which in turn increases CO_2 capture through the longer duration of primary production at higher rates. However, many northern lakes and ponds are already supersaturated with carbon and are net sources to the atmosphere (Jonsson et al., 2003). This source is likely to grow following increases in the supply of organic material from permafrost thawing and carbon transport to the lakes from enhanced terrestrial plant production, but will be determined by the balance between drying and water-logging of the land surface which is as yet unknown.

The processes and feedbacks described above are all biological in nature but physical and chemical processes within the sea ice itself have also been suggested to play a role. Anderson et al. (2004) and Rysgaard et al. (2009) found that levels of dissolved inorganic carbon increase with depth below the surface mixed layer of the Arctic Ocean. This is a result of enhanced air-sea exchange of CO_2 caused by sea-ice formation and the rapid transport of high salinity brine out of the surface layer. Similarly, Alekseev and Nagurny (2007) showed that the polar amplification of the annual cycle of CO_2 over the Arctic Ocean is connected to sea-ice growth, where the formation and growth of ice during the winter is accompanied by an increase of CO_2 in the sub-ice water layer and a CO_2 flux into the atmosphere. Model results showed that melting of sea ice exported from the Arctic Ocean into the East Greenland Current and the Nordic Seas increases the seasonal CO_2 uptake in the area by around 50%. Alekseev and Nagurny (2007) speculated that, overall, the Arctic Basin is more likely to be a source than a sink of CO_2 over an annual cycle.

11.1.2.1.2. Methane (wetlands, lakes, sub-sea)

In anaerobic soils, ponds and lakes of the Arctic, microbes produce CH_4 rather than CO_2. Methane is 25 times more powerful than CO_2 as a greenhouse gas over a 100-year time frame with a radiative forcing of 0.48 W/m^2 (Forster et al., 2007), so any increases in CH_4 emissions will be particularly important. Unlike CO_2, CH_4 is difficult to measure over large areas because there are geographical hotspots for CH_4 production and because there are episodic releases of CH_4. For example, recent research (Mastepanov et al., 2008; Chapter 5, this volume) has shown large releases of CH_4 in the autumn from High Arctic permafrost areas in northeastern Greenland, and thaw lakes in Siberia are hotspots for CH_4 production (Walter et al., 2006, 2007a,b).

In addition, it is difficult to project future releases of CH_4 from land because wetlands on permafrost are drying in some areas (Smith et al., 2005; Riordan et al., 2006; Bunn et al. 2007; Goetz et al., 2007; Chapter 5, this volume), which would decrease current CH_4 emissions. Conversely, thermokarst pond formation is occurring elsewhere with concomitant increases in CH_4 emissions (Christensen et al., 2004; Ström and Christensen, 2007; Myers-Smith et al., 2008; Chapter 5, this volume). A relatively small source of CH_4 can result in positive radiative forcing even though most of a large area can be a net sink for carbon if vegetation is productive.

Particularly large sources of CH_4 occur in former soils that were inundated when sea levels rose after the last ice age. These areas are the vast continental shelves of the Arctic. Recent investigations in the Laptev Sea have shown high levels of CH_4 throughout the water column and extending up to 1800 m in the atmosphere (Shakhova et al., 2008a,b, 2010; Westbrook et al., 2009). Overall, a 1% release of CH_4 stored in subsea hydrates would be equivalent to a doubling of atmospheric CO_2 concentration in terms of its radiative effect. Although the uncertainties of the size of the subsea carbon reserves and their stability are great, the potential risk from CH_4 release, some of which is already being documented (Shakhova et al., 2008 a,b, 2010), is sufficiently large to cause concern.

11.1.2.1.3. Nitrous oxide (tundra peat lands)

Very high emissions of the powerful greenhouse gas nitrous oxide (N_2O; radiative forcing of 0.16 W/m^2, Forster et al., 2007) have recently been discovered from peat circles in patterned ground of the discontinuous permafrost zone in Russia (Repo et al., 2009). These peat circles are widespread in certain regions. Elberling et al. (2010) also reported high N_2O emissions of 34 mg N /m^2/d in cores taken from northeastern Greenland and incubated in the laboratory that are equivalent to daily N_2O emissions from tropical forests on a mean annual basis. These discoveries bring Arctic permafrost N_2O sources, previously thought of as of very little importance, to the agenda for global studies of N_2O dynamics and their impacts via feedbacks.

11.1.2.1.4. Water vapor

Water vapor is a greenhouse gas with a radiative forcing of 0.07 W/m^2 (stratospheric water vapor from CH_4; Forster et al., 2007). In the Arctic, it is transported from the surface to the atmosphere via evaporation from open water and wet surfaces, evapotranspiration from vegetation, and sublimation

from snow (King et al., 2008) and ice. The feedback effects of the processes generating water vapor are complex in that evaporation leads to local cooling, but the increased atmospheric concentrations of water vapor can lead to warming over a wider area because of mixing and transport of water vapor in the atmosphere (positive feedback). It is further complicated by the possibility of increased atmospheric water vapor enhancing cloud formation, which can have either a warming or cooling effect on the surface depending on the cloud height and time of year (Rouse et al., 1997, also discussed below). Assessing and comparing the local versus wider impacts of water vapor will require additional research.

Direct evaporation from melting snow and ice will decrease as the areas of glaciers, lake, pond and river ice, and the duration of ice and snow cover decrease over much of the Arctic (Chapters 3, 4 and 5, this volume). Sublimation should increase initially as the temperature difference between snow and air increases.

Evapotranspiration is expected to increase in those areas of the Arctic where plant production increases, for example in response to a warmer, earlier and longer snow-free season (Euskirchen et al., 2006). Projections of increased forest growth and extent in the Barents region of the Russian Arctic show that by 2080 summer temperatures will have decreased by 1.5 °C due to evaporative cooling (Gottel et al., 2008). Although this cooling is a local effect, the atmospheric mixing of the water vapor is likely to increase warming over larger geographical scales and other feedbacks are also associated with changes in vegetation, such as albedo.

11.1.2.1.5. Ozone and bromine

An important feature of Arctic atmospheric chemistry is the relationship between sea ice, bromine, and ozone. Releases of bromine gas species from the ocean can result in sudden ozone depletion in the lower troposphere during spring. One possible mechanism of bromine release is sea-salt aerosol production from snow lying on sea ice during blowing snow events. A reduction in Arctic sea ice would reduce the importance of this process in the chemistry of the Arctic atmosphere. In chemistry model simulations, Voulgarakis et al. (2009) found large spring ozone increases (up to 50–60%) over the Arctic, due mainly to a reduction in the impact of bromine chemistry, caused by sea-ice retreat. Tropospheric ozone has a relatively small radiative impact (warming), although the effect is greater over bright surfaces (Shindell et al., 2006). With a declining ice cover potentially giving rise to an increase in tropospheric ozone, the feedback would be positive. The impact of a shrinking sea-ice cover on polar stratospheric ozone was studied by Scinocca et al. (2009). They found that the stratospheric response is springtime polar cooling that is dynamical rather than radiative in origin. In the model simulations, the response lags the onset of the sea-ice loss by about a decade. It is associated with an enhanced weakening of the North Atlantic Meridional Overturning Circulation.

11.1.2.2. Transfer of heat and energy between the cryosphere, atmosphere, and ocean

11.1.2.2.1. Albedo

A continued and increased melt of mountain glaciers and ice caps, the Greenland Ice Sheet, and sea ice; an earlier melt of river and lake ice; and a decrease in snow cover will all impact on the Arctic radiation budget. Albedo, the fraction of the incident sunlight that is reflected by the surface, will decrease and more solar radiation will be absorbed by the ground or open water. This will lead to additional warming. Furthermore, a decrease in albedo resulting from a decrease in snow cover over land may result in a further change in albedo through a change in the vegetation in response to surface warming (Chapin et al., 2005; Euskirchen et al., 2007, 2009). Permafrost thaw is also expected to lead to changes in albedo as areas dry out or become waterlogged. Indirectly, changes in surface albedo can modify large-scale circulation (Dethloff et al., 2006).

For the Arctic glaciers (including those in Greenland surrounding the ice sheet) projected volume loss ranges between 0.049 to 0.131 m sea level equivalent (SLE), or 12–32% of their current volume by 2100 depending on the choice of GCM (Radić and Hock, 2010). In the Russian Arctic, there will be differences in the extent of glacial retreat between the Arctic continental climate systems of the

mountains and Kamchatka. Any translation of losses of ice volume into decreased extent will reduce albedo due to increases of bare ground.

Projections of sea-ice extent show a continued and possibly accelerating decline, leading to an ice-free Arctic Ocean in summer within this century and possibly even within the next thirty years. The thinning of the ice pack, earlier melt onset, and increased open water will contribute to a reduction in albedo. While a variety of recent studies have demonstrated the importance of increased absorption of solar radiation resulting from less sea ice (and hence a lower albedo surface), some recent research suggests that the ice–albedo feedback is potentially less efficient than previously thought (Winton, 2006, 2008; Bitz, 2008; Graverson and Wang, 2009). For example, the annual sea-ice minimum is reached in September at a time when incoming solar radiation is already rather weak.

The albedo of terrestrial snow-covered surfaces in the Arctic is a major feedback to climate over a large area (Fernandes et al., 2009), and an important feedback to climate globally through atmospheric and oceanic teleconnections. Recent observations show a general decrease in snow-cover extent and duration that has led to decreases in albedo. Projections indicate increases in snow cover in some areas at the highest latitudes, strong decreases in the southern Arctic, and uncertainty in areas between. These changes will generally lead to an overall net reduction in the albedo feedback. Snow properties also affect albedo, for example, grain size, age of the snow cover and the content of extraneous material such as black carbon from industrial pollution and biomass burning south of the Arctic. In fact, warming resulting from black carbon deposition on snow is equivalent to that resulting from a doubling of atmospheric CO_2 in Eurasia (Flanner et al., 2008; Chapter 4, this volume).

The reflective properties of snow are strongly modified by vegetation in many and complex ways (Pomeroy et al., 2006; Essery et al., 2008; Chapter 4, this volume). While snow typically absorbs around 10% of thermal radiation incident on it, black spruce (*Picea mariana*) can absorb 95% of thermal radiation leading to a significant positive feedback (Juday et al., 2005). The snow feedback will be decreased as vegetation height increases above the snow pack during the ongoing and projected process of shrub and tree range extensions into the Arctic (Sturm et al., 2001; Tape et al., 2006). Although increases in shrub advance in the Alaskan Arctic have not yet resulted in warming, it is predicted that an increase of shrubs could increase summer heating of the atmosphere by 3.7 W/m^2, which is equivalent to a doubling of CO_2 or a 2% increase in the solar constant (Chapin et al., 2005). In the Barents region, changes in albedo through forest advance could increase temperatures by 1 °C in spring. This example should be compared with the cooling effect of increased evapotranspiration in summer and the expected cooling (not yet calculated) of the increased draw-down of atmospheric CO_2 giving an extra 2 to 17 kg C/m^2 by 2080. Together, these projections demonstrate the complexity of feedbacks arising from changes in vegetation: the feedbacks operate in different directions, through different mechanisms and over different periods of time.

11.1.2.2.2 Cloud feedbacks

The exact nature of the cloud radiation feedback in a warming climate is uncertain. Clouds both reflect solar radiation and absorb longwave (terrestrial) radiation, the magnitudes of which depend on cloud amount, height, particle phase and size, and thickness. Some satellite studies have shown that wintertime cloud amount in the Arctic appears to have been generally decreasing since the early 1980s, but springtime cloud amount has been increasing. While wintertime clouds in the Arctic have a warming effect, springtime cloud can have either a net warming or cooling effect. The overall radiative impact of these cloud cover changes on the surface is one of increased cooling. Therefore, changes in cloud cover may have actually suppressed Arctic warming to some degree (Wang and Key, 2003).

The influence of changes in cloud cover on sea-ice extent and *vice versa* is an important part of the feedback process, but has not been studied extensively until recently. On the time scale of a single season, changes in cloud amount may have minimal influence on summer sea-ice melt. The negative cloud anomaly (less than average cloud amount) and increased downwelling shortwave radiation flux from June through August 2007 did not appear to contribute substantially to the record ice extent minimum in September. On the other hand, there are clearly interdependencies between trends in

cloud cover, surface temperature, and sea ice extent. Over the past few decades, more than 80% of the observed surface warming in the western Arctic Ocean during autumn is attributable to decreasing sea ice. Similarly, over 80% of the winter surface cooling in the central Arctic is a result of changes in cloud cover. In spring, only about half of the surface warming is a result of changes in cloud cover (Liu et al., 2009).

Satellite and reanalysis data have shown that sea-ice retreat is linked to a decrease in low-level cloud amount and an increase in mid-level clouds (Schweiger et al., 2008). This is in contrast to the common notion that a warming ocean surface will increase surface evaporation and lead to more low clouds. While surface evaporation does increase, near surface atmospheric instability increases, so cloud formation occurs in the mid-troposphere rather than nearer the surface. However, the radiative effect of these changes is relatively small, as the cloud radiative forcing changes are compensated for by changes in the lower tropospheric temperature and humidity structure. Because of the seasonal timing of this process, the response of cloud cover to sea-ice loss plays a minor role in regulating the summertime ice-albedo feedback. Its role in the cloud-ice feedback in autumn is potentially larger, although the ice-albedo feedback itself is weaker in autumn due to the relatively low amount of solar radiation at very high latitudes (Kay and Gettleman, 2009).

Increasing temperatures and levels of ultraviolet (UV)-B radiation affect biological emissions from marine ecosystems (Zepp et al., 1995). The grazing of phytoplankton by zooplankton triggers the release of dimethyl sulfide (DMS), which is emitted to the atmosphere where it acts as condensation nuclei for water droplet formation. In other words, the release of DMS from the surface stimulates cloud formation. Furthermore, UV-B radiation converts some dissolved organic sulfur to carbonyl sulfide and warming of the ocean reduces its solubility, thereby releasing more to the atmosphere where more cloud nuclei are formed. The currently increasing presence of sea-ice leads during the summer melt in the Arctic Ocean increases the biological activity of the region. This may result in an accumulation of organic material, especially in the surface microlayer. Particulate organic matter, including microorganisms, small water-insoluble particles and microaggregates, can form a substantial part of the summer aerosol over the open leads in the central Arctic Ocean, thereby enhancing cloud formation (Matrai et al., 2008).

11.1.2.2.3. Atmospheric circulation

Overall, the atmosphere is a driver of change in the Arctic (Francis et al., 2009). It primarily forces rather than responds to changes in the cryosphere, a good example being changes in atmospheric wind patterns over the past decade that have contributed to recent reductions in summer Arctic sea-ice extent (Ogi et al., 2010). But the increase in late summer open water area has, in turn, directly contributed to a modification of large-scale atmospheric circulation patterns. With a reduction in sea-ice cover in late summer, additional heat that was stored in the ocean is then released to the atmosphere in autumn. The transfer of heat from the surface to the troposphere is more effective as the strength of the normally stable Arctic boundary layer is eroded (Overland and Wang, 2010). In years with reduced sea-ice cover the lower tropospheric thickness is greater. This has a large-scale impact, even into the northern mid-latitudes, as the pressure fields and therefore winds are directly related to the atmospheric thickness. Responses can be complex, as the loss of sea ice north of Eurasia may result in a cooling effect over eastern Asia.

There is also a relationship between snow cover and atmospheric circulation. Recent observational studies show that above-normal winter snow depth over European Russia and a corresponding below-normal snow depth over central Siberia – the east-west snow dipole – are associated with reduced Indian monsoon rainfall and above normal sea-surface temperatures (SSTs) over the eastern and central tropical Pacific Ocean during subsequent winters (Ye and Bao, 2001; Ueda et al., 2003; Peings and Douville, 2010). Similarly, below-normal winter snow depth over European Russia and a corresponding above-normal snow depth over central Siberia are associated with increased monsoon rainfall and below normal SSTs. Some GCMs also show this relationship, but the patterns may differ somewhat from the observations. Forcing from the Himalaya region dominates, where a delayed snowmelt leads to a reduction in surface sensible heating and consequently a weakened latitudinal

temperature gradient. The result is less Indian rainfall, particularly during the early part of the summer season.

The large-scale effect of changes in snow cover through the snow albedo feedback (SAF) has recently been examined. For example, Fletcher et al. (2009) demonstrated a non-local influence of SAF on the summertime circulation in the extratropical Northern Hemisphere. In models with stronger SAF, increased land surface warming is associated with large-scale sea level pressure anomalies over the northern oceans and a poleward intensified subtropical jet. This would result in a change in heat and moisture fluxes into/out of the Arctic, although the feedback on snow cover and the impact on sea ice is not clear.

Accumulation and ablation on ice sheets, ice caps, and glaciers depend on the atmosphere circulation patterns, and atmospheric circulation is, in turn, affected by the presence of, and changes in, these features. Ice sheets and ice caps change the elevation and form of the Earth's surface. This changes the air temperature and deforms the atmospheric circulation, forming a 'wave shadow'. For this reason, the Icelandic Low exists throughout the entire year, with the Greenland Ice Sheet to the west, as opposed to the Aleutian Low, which disappears in summer. Smaller ice caps influence the atmosphere on smaller scales. Due to the cooling effect of the Greenland Ice Sheet, the air temperature drops as much as 5 to 10 °C in the atmospheric layer several hundred meters above it. Over the ice caps (Franz Josef Land, Svalbard, Novaya Zemlya, Severnaya Zemlya) and large mountain glaciers, the cooling is about 2 to 3 °C and spreads upward to 200 to 250 m. This may affect the cyclone trajectories ('storm tracks') and the life cycle of pressure systems (Krenke, 1982). Katabatic winds over the glacier surface have an impact on near surface atmospheric circulation. The intensity of such processes depends on changes in glacier size.

Changes in the patterns of atmospheric circulation due to glacier and ice cap retreat and/or disappearance may change the accumulation and ablation rates on the glaciers. For example, more cyclones in winter leads to more accumulation; more anticyclones in summer leads to more ablation. However, the consequences are currently unknown.

Recent modeling by Lunt et al. (2004) supports the influence of the Greenland Ice Sheet on atmospheric circulation. In their study, the conceptual removal of the Greenland Ice Sheet results in an increase in atmospheric thickness (geopotential height) in that region. The region of cooling in the Barents Sea in December, January and February is most intense at the surface and is associated with substantial growth of the sea ice throughout the year. The cooling appears to be forced from above by the atmosphere, and amplified by the sea ice-albedo effect. The cooling appears to be linked with enhanced equatorward flow east of Greenland, associated with the increased geopotential height over Greenland, which brings cold air from the pole equatorward over the Barents Sea. In addition to changes in the stationary waves in the atmosphere, the conceptual removal of Greenland also affects the transient behavior of the atmosphere. In the no-Greenland case, the intensity of the North Atlantic winter storm tracks is decreased. Associated with this is a corresponding decrease in large-scale precipitation. The decrease in storm track activity is likely to be related to the fact that in the no-Greenland case, the meridional temperature gradient is decreased over the North Atlantic.

A retreating sea-ice margin may enhance melting over the Greenland Ice Sheet. Rennermalm et al. (2009) explored the spatial and temporal covariance of sea-ice extent and ice sheet surface-melt around Greenland from 1979 to 2007. Significant covariance was found in western Greenland. An examination of wind direction patterns and a lag analysis of ice retreat/advance and surface-melt event timings suggested that a change in sea-ice extent is a potential driver of ice-sheet melt, in that late summer wind directions bring onshore advection of ocean heat, enhanced by reductions in offshore sea ice.

There is a strong linkage between sea-ice loss and terrestrial permafrost temperature. Lawrence et al. (2008) examined how rapid sea-ice loss affects the terrestrial Arctic climate and ground thermal state using a GCM. It was found that the accelerated warming signal penetrates up to 1500 km inland and substantially increases ground heat accumulation. Furthermore, enhanced heat accumulation leads to rapid degradation of warm permafrost and may increase the vulnerability of colder permafrost to

degradation under continued warming. In contrast to the relationship between snow cover and the Indian monsoon, which operates on a yearly time scale, the changes in atmospheric circulation impact permafrost on decadal scales.

11.1.2.3. Interactions with the freshwater budget of the Arctic

All of the cryospheric components play, to varying degrees, roles in the freshwater budget of the Arctic. Changes in the cryosphere can affect the strength of the thermohaline circulation in the North Atlantic, where it is called the Atlantic Meridional Overturning Circulation (AMOC), and hence, global climate. Assessing the magnitude of these effects requires not only an understanding of how the thermohaline circulation responds to freshwater inputs, but also the sources, locations, distributions, and pathways of freshwater into and out of the Arctic (Randall et al., 2007).

In general, the Arctic Ocean receives freshwater inputs from direct precipitation, Pacific water via the Bering Strait (referenced to a particular salinity), terrestrial ice masses, and runoff from river basins, which includes large areas of the Northern Hemisphere continents. Importantly, the major river basins contributing flow to the Arctic Ocean are of a nival regime (runoff is dominated by snowmelt), where the snowmelt freshet is the largest and most important hydrological event of the year. Within the Arctic Ocean, freshwater amounts also change due to losses from evaporation and to the growth (-) and ablation (+) of sea ice. Very large volumes of freshwater can also be 'stored' in deep basins with highly variable residence times. Freshwater export from the Arctic Ocean occurs primarily through Fram Strait and the Canadian Archipelago to the Atlantic Ocean, where it plays a role in the formation of deep water in the Greenland-Iceland-Norwegian (GIN) Seas.

It is the export of freshwater that has been identified within paleo-records as weakening the thermohaline circulation and causing major cooling events over the North Atlantic. For example, an Arctic Ocean pathway fed by freshwater flows from the Mackenzie River-Beaufort Gyre, instead of the previously supposed Great Lakes-St. Lawrence routing, has been recently identified as being responsible for the shutting down of the North Atlantic thermohaline circulation, resulting in major cooling associated with the Younger Dryas (Peltier, 2007; Murton et al., 2010).

Since publication of the Arctic Climate Impact Assessment in 2005 (ACIA, 2005), a number of updates have been made concerning the relative size of the overall freshwater budget terms and of the cryospheric components that contribute to them. These are considered in Section 11.1.2.3.2, gleaned from the preceding chapters of this report and related literature. First, some example post-ACIA freshwater budget evaluations are considered.

11.1.2.3.1. Recent budget estimates

Total freshwater inputs to the Arctic Ocean, calculated at about 8500 km^3/y, are dominated by river flow (38%), inflow through the Bering Strait (30%), and precipitation-evaporation directly occurring on the Arctic Ocean (24%) (Serreze et al., 2006). Importantly, this estimate is an order of magnitude smaller than the total amount stored in the Arctic Ocean. Freshwater exports from the Arctic Ocean occur principally through the straits of the Canadian Arctic Archipelago (35%) and via Fram Strait as liquid (26%) and sea ice (25%). For such calculations, the volumes of freshwater are referenced to a mean ocean salinity of 34.8. Also recognizing that a steady state of the freshwater budget should not be expected, Serreze et al. (2006) noted that their values indicate larger freshwater inflow through Bering Strait and larger liquid freshwater outflow through Fram Strait than earlier estimates by others.

Peterson et al. (2006) conducted an analysis of changes in freshwater budget components for a broader 'Arctic region' than Serreze et al. (2006), which included the additional large land-ocean catchment of Hudson Bay in North America as well as the Nordic Seas and North Atlantic subpolar basins. Increasing precipitation-evaporation over the marine environments and larger river flow, probably also tied to increases in high-latitude precipitation, was estimated to have contributed ~20 000 km^3 of freshwater to the total region from lows in the 1960s to highs in the 1990s. Notably, the river trend included a decline in flow for the Hudson Bay system. Sea-ice ablation added a slightly smaller amount of ~15 000 km^3, and glacial melt added ~2000 km^3. Due to the lack of complete mass-balance

estimates for the Greenland Ice Sheet, its contributions were excluded from the latter amount but it was noted to have been ~80 km^3/y in the 1990s and to have recently increased to ~220 km^3/y (Peterson et al., 2006). The most recent estimate of current annual loss of ice from the Greenland Ice Sheet is similar at about 205 ± 50 Gt/y[1] (2005–2006) (Chapter 8, this volume).

11.1.2.3.2. Freshwater budget components and changes

H1_Snowmelt and river discharge

Compared to all other world oceans, the Arctic Ocean receives a disproportionately large amount of river runoff to its total volume via the Lena, Mackenzie, Ob, and Yenisey rivers that are dominantly nival rivers. River flow provides the largest input to the Arctic Ocean freshwater budget (Prowse and Flegg, 2000). Although the seasonality of observed multi-decadal increases in total Eurasian river flow to the Arctic Ocean has not been fully examined, Chapters 4 and 5 (this volume) note that some increases in the magnitude and advances in timing of the snowmelt freshet on northern rivers have been observed and greater changes are expected in the future. This will be due not only to changes in the size and seasonality of snowmelt but also to the effects of thawing permafrost in changing flow pathways and storage. Such changes in the timing and magnitude of flows are important to how river water is distributed, directed and/or stored in the Arctic Ocean (e.g., Cooper et al., 2008; Jones et al., 2008).

H1_Small mountain glaciers and ice caps, and the Greenland Ice Sheet

The overall freshwater contribution from the glaciers (328,556 km^2) in a broadly defined, pan-Arctic drainage basin, as well as small ice caps around the Greenland Ice Sheet (but not including the ice sheet itself), is ~1700 km^3 for the period 1961 to 2001 (Dyurgerov and Carter, 2004). Freshwater contributions from these glaciers varied annually from near zero to just under 200 km^3, far less than contributed by the corresponding nine major rivers (5282 km^3/y) within the same pan-Arctic region. However, there is a greater 'positive change signal' from the glaciers than river discharge (Dyurgerov and Carter, 2004). Partly included in this total are the glaciers from the Alaska-Yukon region, which contribute to the North Pacific Ocean waters near the principal Arctic Ocean inflow point, the Bering Strait. These glaciers have experienced some of the most rapid wastage. In the case of Alaska, for example, freshwater contributions have increased from 52 ± 15 km^3/y over the 1950s to 1990s period to ~96 km^3/y from the mid-1990s to 2001 (Arendt et al., 2002), and more recent sampling suggests increased ablation particularly at low elevations. In reference to the Bering Strait to which these glaciers contribute, flow volumes to the Arctic Ocean have been significantly revised upward (i.e., by ~50% to ~2500 km^3/y; Woodgate and Aagaard, 2005), although the exact contribution of ablating glaciers to such volume increases remains unknown.

Some freshwater budget analyses contain little discussion about the role of the Greenland Ice Sheet, despite its strategic placement as a freshwater source in the North Atlantic. In addition to freshwater volume, the location of the input may also be important (Randall et al., 2007) and meltwater runoff from the ice sheet is potentially a major source of freshening that has not yet been included in relevant models (Randall et al., 2007).

H1_Sea ice

The Arctic Ocean is a salt- rather than temperature- stratified ocean and hence, sea ice growth/ablation and ocean dynamics can be greatly modified by changes in freshwater. It is the salt-stratified upper layers that provide the vertical stability to permit formation of an ice cover. Surface freshwater layers, however, also contribute with the halocline to thermally shield sea ice from bottom melt, which can be driven by the large quantities of heat contained in warmer, deeper water originating from the Atlantic Ocean. Carmack (2000) estimated the volume of ice forming and melting each year amounted to a 1.45 m equivalent of freshwater (depth of freshwater across the ocean) assuming an un-deformed ice

[1] [Add conversion factor]

cover, and an additional 0.45 to 0.7 m of freshwater if the mass in ice ridges is also considered. However, sea ice has undergone significant changes in areal coverage and thickness.

For some key episodic losses, sea ice-bottom melt has been linked to solar heating of the upper ocean (Perovich et al., 2008). Bottom melt can also result from the loss of thermal insulation from the warmer Atlantic water provided by the surface layers of freshwater and cold halocline. The stability of these upper layers, particularly with enhanced vertical mixing, has been identified as a 'key wild card' regarding future sea-ice loss (Serreze et al., 2007). Sea ice is also exported through Fram Strait along with sea ice meltwater, but export of sea ice meltwater seems to be the least likely to influence thermohaline circulation (Jones et al., 2008). This could play a large role in changing ice conditions.

H1_Ocean storage and pathways

The fate of sea ice meltwater and other forms of freshwater is not simply direct export because the Arctic Ocean also holds significant freshwater in storage with variable releases. While about one-quarter of the total is held on shelves, the majority is in the Eurasian and Canada basins, the latter being the largest single freshwater storehouse in the Arctic Ocean. This has been ascribed to the deep halocline of Canada Basin, which stores freshwater from sea ice meltwater, meteoric water (ocean precipitation and terrestrial runoff), and low-salinity Pacific water from the Bering Strait. Estimates of this storage (like the other freshwater budget terms) vary in the literature, and are primarily due to changes in import-export to Canada Basin and, in the accuracy and ability to measure its content. Despite variations in its estimated volume (e.g., ~46 000 km^3, Carmack, 2000) and 25 600 km^3 (Yamamoto-Kawai et al., 2008), it is generally accepted that the largest source of the average annual freshwater input (~3200 km^3) to this and other freshwater storehouses in the Arctic Ocean (Yamamoto-Kawai et al., 2009) is river runoff – estimated by Yamamoto-Kawai et al. (2008) for Canada Basin to be 800 km^3/y and to be just slightly smaller than the amount removed by sea ice formation (900 km^3/y). They further estimate that the average export of ice and liquid freshwater from Canada Basin contributes ~40% of the freshwater flux to the North Atlantic.

The storage values vary with time due to atmospheric circulation, which can control pathways of freshwater to/from storage basins as well as the storage/release from within the storage basins (i.e., Ekman pumping, which under anticyclonic [cyclonic] circulation stores [releases] freshwater). Most recently, measurements from Canada and Makarov basins indicate that there has been a freshwater storage increase of up to 8500 km^3, and by extrapolation, almost 11 000 km^3 in all the deep basins of the western Arctic. By contrast, the Eurasian Basin in the eastern Arctic and closer to the main export to the North Atlantic has experienced a loss of about 3300 km^3, giving a net gain of 7700 km^3 (McPhee et al., 2009). This is a significant increase being approximately four times the volume associated with the Great Salinity Anomaly (a near-surface pool of fresher-than-usual water tracked in the subpolar gyre currents from around 1968 to 1982, which affected regional climate) and similar in magnitude to the total 1981 to 1995 sea ice attrition (melt plus export) estimated in the above noted freshwater budget by Peterson et al. (2006).

11.1.2.3.3. Model projections

At the time of the Arctic Climate Impact Assessment (ACIA, 2005), there was a concern about the intensification of the hydrological cycle at high latitudes and the effect this would have on the AMOC. ACIA noted that projections by atmosphere-ocean GCMs produced varying results with changes in the maximum strength of the AMOC by the end of the 21st century ranging from zero to a reduction of 30–50%. The suite of ACIA-selected GCMs (Kattsov and Källén, 2005) did not include freshwater runoff from melting ice sheets and glaciers. Hence, it was surmised that the model AMOC sensitivity to global climate change might be too weak, although sensitivity experiments indicate freshwater runoff must be several times 'present-day' volumes to appreciably alter the AMOC.

A strong scientific debate remains about the potential significance of freshwater effects on the AMOC (Randall et al., 2007). For example, Holland et al. (2007) noted that a constituent result among models for the period 1950 to 2050 (observations and modeled results) is an acceleration of the hydrological cycle, including increased ocean net-precipitation, river runoff, and net sea-ice melt. They also noted,

for liquid water, a larger export to lower latitudes, primarily through Fram Strait, and storage in the Arctic Ocean. By contrast, export and storage of freshwater in the form of sea ice decreases, although there is significant variability in sea-ice budget terms. Largely similar results were reported by Koenigk et al. (2007).

A number of efforts are underway to more accurately define the role of freshwater in AMOC weakening, the Coupled Model Intercomparison Project (CMIP) and Paleoclimate Modelling Intercomparison Project (PMIP) coordinated effort being major examples. All models used in CMIP show that AMOC weakening projected for the 21st century is caused more by changes in the surface heat flux than by freshwater, although its effect on high-latitude stratification plays a contributing role (Arzel et al., 2008). They further noted that interannual exchanges in freshwater between the GIN Seas and the North Atlantic have a major driving influence on the interannual variability of deep convection over the 21st century.

11.1.3. Synthesis

This review and analysis has substantiated and quantified many of the linkages between the cryosphere and climate identified by earlier recent studies (e.g., Francis et al., 2009). Figure 11.2 synthesizes the major feedbacks on climate mediated by cryospheric processes, where the impacts of changes in various forcing variables (e.g., greenhouse gases, albedo) and processes (e.g., atmospheric circulation and thermohaline circulation) on elements of the cryosphere are depicted. The surface or near-surface air temperature is the variable primarily used to drive changes in the cryospheric component and then to assess the magnitude of warming or cooling. The box colors indicate the expected future impact of a change in the variable or process on the climate, where red indicates warming and blue indicates cooling. The colors also indicate the type of feedback, where red is a positive feedback from current climate warming to future temperature and blue is a negative feedback. The color does not, however, indicate the direction of change between the two variables alone. For example, a decrease in sea-ice cover increases atmospheric water vapor, which in turn may decrease sea-ice cover even further because water vapor is a greenhouse gas. This is a positive feedback, but the change in sea-ice cover and water vapor are in opposite directions. The intensity of the colors indicates, at least qualitatively, the relative magnitude of the impact. For example, warming caused by a decrease in albedo through a loss of sea ice is expected to be greater than that due to changes in glacier extent. The feedbacks presented in Figure 11.2 represent a relatively simple synthesis of many complex processes. This synthesis is intended to stimulate a better understanding of the processes rather than give a definitive analysis.

Figure 11.2. The major feedbacks on the climate via the cryosphere. The box colors indicate the expected future impact of a change in the variable or process on the cryosphere (where red indicates warming and blue indicates cooling). The intensity of the colors indicates, at least qualitatively, the relative magnitude of the impact or levels of understanding (where grey represents uncertainties due to conflicting information etc. and white indicates a knowledge gap). The change in intensity of colors from left to right indicates an acceleration or retardation of the process assuming a time horizon over ~30 yrs. The feedbacks presented in this figure represent a relatively simple synthesis of many complex processes.

The impacts, interactions, and feedbacks in Figure 11.2 are not necessarily independent. In particular, cloud formation is strongly dependent upon the available water vapor, which may be from local sources or transported from lower latitudes (atmospheric circulation). Additionally, changes in one part of the cryosphere may affect changes in others. For example, thin ice has a very high growth rate but is strongly affected by snow cover. Snow is an efficient insulator, so ice with little snow on its surface can grow much faster than ice with a thick snow cover. This allows for the newly forming sea ice to reach a thickness that might even be larger than that of ice that survived the summer. Although the ice-albedo feedback leads to a decreasing summer sea-ice cover, rapidly growing thin ice leads to a recovery of the sea-ice cover during winter. An additional contributor to this recovery is that often there is little snow on the newly forming sea ice. It is presently unclear how an ice-diminished Arctic will impact on snowfall, so it is difficult to assess to what degree the growth of thin ice may offset the increase in ice melt through the ice-albedo feedback.

Despite the interactions between feedbacks, Figure 11.2 provides a snapshot of how changes in the climate system affect the cryosphere, illustrating the signs and magnitudes of the feedbacks.

11.1.4. Uncertainties and recommendations

Most of the terrestrial feedbacks are poorly quantified over large areas although considerable detail is available for a few specific locations. Generalization of the impacts is difficult at best. *A better observational capacity is needed in terms of geographical extent and prolonged periods of observation while it is necessary to reconcile effects of thawing permafrost and changing snow cover on tundra drying versus wetting as this is a major determinant of terrestrial wetland trace gas emissions.*

Large uncertainties exist in cloud feedbacks and subsea permafrost: *these topics require enhanced and sustained research.*

General circulation models currently include few feedbacks. *Modeling studies including sensitivity analyses need to be intensified to include all the major feedbacks and to calculate the net effects.*

Lack of full coupling between surface dynamics and the atmosphere is a major gap in current GCMs. Chapter 8 (this volume) *recommends more accurate incorporation of surface albedo and snow microphysics in climate models.*

There remains considerable controversy about the degree to which current levels of freshwater within the Arctic Ocean can affect the strength of the AMOC. *There is a need to evaluate the potential for the cumulative production and release of large amounts of freshwater from all contributing components, including all components of the cryosphere. Also there is a need to assess whether there are specific circulation patterns and durations that could cause significantly large freshwater production from all terrestrial and marine freshwater sources and their ultimate export to the North Atlantic where they could produce a significant effect on the AMOC and global climate.*

The primary source of uncertainty regarding the atmosphere as a driver of change in the Arctic is precipitation. Future changes in its spatial and seasonal distribution are unclear, as are its effects on the sea-ice mass budget, marine primary productivity, and vegetation. *A better understanding of precipitation effects will only be gained through developing and sustaining a more robust observing network.*

Acknowledgements

We thank Sebastian Gerland, Mats Granskog and two reviewers for comments on the manuscript. Terry Callaghan and Margareta Johansson gratefully acknowledge the support from the Swedish Research Councils FORMAS and the Environmental Protection Agency for supporting their involvement in the SWIPA process.

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11.2. Sea-Level Change

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11.2.1. Introduction

This discussion of sea-level change is framed in terms of the impact on global mean sea level of net loss of mass from Arctic mountain glaciers and ice caps, and the Greenland Ice Sheet. Reconstructions of sea-level change using geological evidence show that a sea-level rise of one or more metres per century was not uncommon in the past. It has recently been suggested that such rates occurred during the last interglacial warm period 120 000 years ago when the volume of land ice was similar to that at present (Berger, 2008; Rohling et al., 2008). Worldwide, the coastal zone changed profoundly during the 20th century, primarily due to a growing population and increasing urbanization. In 1990, 23% of the world's population (or 1.2 billion people) lived within 100 km distance and 100 m elevation of the coast at densities about three times higher than the global average (National Research Council, 2011). Around 200 million people live on coastal floodplains, less than 1 m above current sea level. Sea-level rise is thus one of the major global socio-economic hazards associated with climate change. With coastal development continuing at a rapid pace, society is becoming increasingly vulnerable to sea-level rise and variability. Rising sea levels will contribute to more severe storm surges and flooding, even if hurricane intensities do not increase in response to the warming of the oceans. They will also contribute to the erosion of the world's sandy beaches, most of which have been retreating over the past century. Low-lying islands are also vulnerable to sea-level rise.

Box 11.1. Sea-level terminology

Mean sea level (MSL) is defined as the average height of the ocean surface (i.e., the global average of the elevations of the mid-points between local mean high and low tide levels). The term *eustatic* refers to changes in global mean sea level due to changes in the mass of water in the oceans, but also through tectonic processes which change the size or shape of the ocean basins. The term *steric* refers to changes in global mean sea level due to changes in water density arising from thermal expansion or compression and/or salinity variations. The *local mean sea-level* (LMSL) refers to the height of the sea with respect to a local benchmark on land, and is determined by averaging the height difference over a period of time long enough to smooth out fluctuations caused by waves and tides. In some regions, vertical movements of the land can be of the same order (mm/y) as sea-level changes, and measurements of local mean sea level must be corrected for this effect. Such land movements can occur because of ongoing *isostatic* adjustment of the mantle to temporal variations in surface loading resulting from changes in glacier or ice sheet mass. Isostatic adjustments are greatest near the sources of glacier or ice sheet change (where they are referred to as *glacio-isostatic* adjustments), but combine with other load signals to form a complex global pattern of crustal uplift and depression (National Research Council, 2011).

11.2.2. Instrumental sea-level record

Since 1870, global sea level has risen by about 0.2 m (Bindoff et al., 2007). During the 20th century, the rate of rise in global mean sea level was in the range 1.0 to 2.0 mm/y (Church et al., 2001), with recent analyses for the past 50 years giving values in the upper part of this range (Holgate and Woodworth, 2004; Church and White, 2006). During this period, the ice sheets of Antarctica and Greenland made a relatively small contribution to sea-level change (a few tenths of a millimetre per year at most; Rignot and Thomas, 2002), so the changes observed were due primarily to the combination of steric changes and global mountain glacier / ice cap melt. Since 1993, sea level has been measured accurately globally using altimeters on satellite platforms. Before then, the data were obtained from tide gauges at coastal stations around the world. Satellite and tide-gauge measurements both show that the rate of sea-level rise has accelerated since 1990 (Figure 11.3). Satellite measurements show that sea level rose at an average rate of 3.3 ± 0.6 mm/y from 1993 to 2008 (Ablain et al., 2009). This decreased to 2.5 ± 0.6 mm/y for the period

2003 to 2008 (Ablain et al., 2009; Cazenave and Llovel, 2010). Statistical analysis reveals that the rate of rise is correlated with global mean surface temperature (Rahmstorf, 2007; Grinsted et al., 2009). Sea-level rise is a predictable consequence of climate warming for two main reasons: ocean water expands as it heats up, and additional water flows into the oceans from the ice that melts on land.

11.2.3. The role of Arctic mountain glaciers and ice caps, and the Greenland Ice Sheet

The chapters addressing Arctic mountain glaciers and ice caps (Chapter 7, this volume) and the Greenland Ice Sheet (Chapter 8, this volume) document present knowledge of historical changes in the mass of these ice bodies. Net mass loss from Arctic mountain glaciers and ice caps has increased and corresponds to a rise in mean sea level of 0.5 mm/y in the most recent decade (ca. 1996 to 2006, Dyurgerov et al., 2009) and perhaps even to 0.75 mm/y in the period 2000 to 2005 (Cogley, 2009). The volume of all the world's mountain glaciers and ice caps (all land ice except the Greenland and Antarctic ice sheets) is estimated to be equivalent to a change in mean sea level of 0.60 ± 0.07 m; the Arctic mountain glaciers and ice caps contain 68% of the total volume of these glaciers and ice caps which is equivalent to 0.41 m of mean sea-level rise (Radić and Hock, 2010). The rate of mass loss from the Greenland Ice Sheet has been increasing rapidly, and its contribution to mean sea level rise has progressively increased from about 0.14 mm/y (1995–2000), to 0.28 mm/y (2000–2005), to 0.57 mm/y (2005–2006) (averages from Figure 8.??, this volume, assuming 1 mm/y = 360 Gt/y). The most recent estimate of the total mass loss from the Arctic mountain glaciers and ice caps, and the Greenland Ice Sheet (for 2003–2008) thus sums to 1.3 mm/y mean sea-level rise (462 Gt/y) and hence contributes more than half to the most recent estimate of the global rate of mean sea-level rise (2.5 mm/y; Ablain et al., 2009). Both the proportional contribution from the Greenland Ice Sheet and the total contribution from Arctic land ice (i.e., the Greenland Ice Sheet, and the Arctic mountain glaciers and ice caps) to mean sea-level rise have increased strongly since 1995 (Figure 11.4). The uncertainties on the sea-level rise estimates for the Greenland Ice Sheet and the Arctic mountain glaciers and ice caps vary within the literature, with values of 0.1 to 0.2 mm/y on each of the components resulting in an uncertainty in the sum of the estimated sea-level rise from the two sources combined of 0.23 mm/y (Ablain et al., 2009 Cazenave and Llovel, 2010).

11.2.4. Additional sources contributing to sea-level rise

In order to assess the causes of observed sea-level changes and predicted future sea-level rises, the contributions from steric sea-level rise need to be included to complete the picture; consisting mainly of thermal expansion of the ocean water, mass losses from the Antarctic Ice Sheet and non-Arctic mountain glaciers and ice caps, and change in land water storage. The loss of mass of the global glaciers and ice caps has increased from 0.8 ± 0.1 mm/y (1993–2003) to 1.4 ± 0.25 mm/y (2003–2007) (Cazenave et al., 2008; Cazenave and Llovel, 2010). The Antarctic Ice Sheet is the largest body of ice in the world, and uncertainties in glacial isostatic adjustments in this area are large, making it difficult to obtain accurate estimates of ice sheet mass loss. The most reliable estimates have been determined from satellite observations made during the past decade. The loss of mass from the Antarctic Ice Sheet is observed to be equivalent to 0.2 ± 0.17 mm/y mean sea-level rise in 1993–2003, increasing to 0.5 ± 0.15 mm/y in 2003–2007 (Cazenave et al., 2008; Cazenave and Llovel, 2010). Thermal expansion of the ocean water was the greatest component of observed mean sea-level rise with a value of 1.0 ± 0.3 mm/y in 1993–2003 (Cazenave and Llovel, 2010). In the most recent period the contribution from thermal expansion has decreased to 0.25 ± 0.8mm/y (Cazenave and Llovel, 2010). The rate of thermal expansion is believed to be closely related to the changing global mean surface temperature. The final component of the mean sea-level rise is the change in storage of water on land due to climatic change and human activities. Estimates based on GRACE satellite observations suggest a negative change in terrestrial water storage in the period 2003–2007, resulting in a mean sea-level rise of -0.2 ± 0.1 mm/y (Cazenave and Llovel, 2010). A compilation of historical contributions from thermal expansion, the different land ice components, and change in terrestrial water storage is shown in Figure 11.5 and Table 11.1.

Table 11.1. Contributions to mean sea-level rise for the periods 1993–2007 and 2003–2007. Quoted errors are one standard deviation. Source: Cazenave and Llovel (2010).

Sea-level rise, mm/y	1993–2007	2003–2007
Observed	3.3 ± 0.4[a]	2.5 ± 0.4[a,b]
Thermal expansion	1.0 ± 0.3[c]	0.25 ± 0.8[d] (Argo)
Ocean mass	2.3 ± 0.5[e]	2.1 ± 0.1[f]
Mountain glaciers and ice caps	1.1 ± 0.25[g]	1.4 ± 0.25[h]
Total ice sheets (Greenland and Antarctic)[j]	0.7 ± 0.2	1.0 ± 0.2
	(0.4 ± 0.15)	(0.5 ± 0.15)
	(0.3 ± 0.15)	(0.5 ± 0.15)
Land waters	–	-0.2 ± 0.1[k]
Sum of (2 + 4 + 5 + 6)	2.85 ± 0.35	2.45 ± 0.85
Observed rate minus sum	0.45	-0.05

[a] The observed sea-level rise is GIA corrected (-0.3 mm/y removed); [b] Ablain et al. (2009); [c] mean of Levitus et al (2009) and Ishii and Kimoto (2009) values; [d] mean of Willis et al. (2008), Cazenave et al. (2009) and Leuliette and Miller (2009) values; [e] observed rate minus thermal expansion; [f] GRACE with a -2 mm/y GIA correction, Cazenave et al. (2009); [g] based on Kaser et al. (2006) and Meier et al. (2007); [h] Cogley (2009); [j] compilation of published results; [k] W. Llovel, K. DoMinh, A. Cazenave, J,F. Cretaux, M. Becker, unpublished manuscript.

11.2.5. Projections into the future

Radić and Hock (2011) computed estimates of the mass changes of more than 120 000 mountain glaciers and around 2600 ice caps around the world to 2100 using an elevation-dependent temperature-index mass balance model driven by output from ten general circulation models (GCMs) forced by the IPCC A1B emissions scenario. Future volume changes were up-scaled to all glaciers using a regionally differentiated approach. For the Arctic mountain glaciers and ice caps (including those in Greenland surrounding the ice sheet) the projected volume losses due to melt (mass loss by calving is not included) show a large range (51–136 mm mean sea-level equivalent (SLE) or a 13–36% reduction in their current volume by 2100) depending on the choice of GCM (Figure 7.23, this volume). The modeled volume reduction by 2100 varies considerably among regions, with the smallest reductions in Greenland (8 ± 4 %) and the largest in Svalbard (54 ± 15 %), with a large spread among models for most regions (Figure 7.23, this volume). Most models show the greatest Arctic sea-level contributions coming from the Canadian Arctic, Alaska (including the glaciers of northwestern Canada), followed by Svalbard and the Russian Arctic. Contributions from Arctic Canada show a very large range; from a mean sea-level equivalent contribution of <10 mm to the largest projected contribution for any of seven Arctic regions, indicating a large spread in the projected temperature and precipitation fields among the GCMs for this region (Figure 7.24, this volume).

The above projections do not include mass losses due to calving that result from changes in the dynamics of tidewater glaciers. In the polar regions, there is evidence of accelerating flow and increased calving of tidewater glaciers, especially in Greenland and in western Antarctica, but it is not yet possible to predict reliably whether this will become more widespread, or for how long it may be sustained.

Since the IPCC Fourth Assessment, a number of new studies of ice-sheet mass budget have considerably enhanced understanding of the potential for rapid collapse of ice sheets in response to changing climate (Allison et al., 2009). Recent observations have shown that changes in the rate of ice discharge into the ocean can occur far more rapidly than previously suspected (e.g., Rignot, 2006). Currently there is no dynamic ice-sheet model that can predict the response of the Greenland Ice Sheet to a warmer climate on a century time scale. To get around this difficulty, Pfeffer et al. (2008) attempted to assess the rates of sea-level rise that could occur under a range of physically plausible scenarios. Under their 'Low 1' scenario, the sea-level rise contribution from Greenland is estimated by assuming a doubling in ice discharge, and a continued increase in losses by surface melt from both the ice sheet and mountain glaciers and ice caps at

present-day rates. In this case, the total contribution to sea-level rise would be about 0.17 m by 2100, with contributions of 0.09 m from ice discharge, and 0.07 m from surface melt, respectively.

In western Antarctica, the Amundsen Coast basin (which includes Pine Island and Thwaites Glaciers), has shown increased mass loss in recent years. These marine-based glaciers contain about 1.5 m sea-level equivalent of ice, and are particularly sensitive to changes in climate and in ocean circulation and temperature. The average ice velocity in this region is 2 km/y, which is considerably higher than the average velocity for all Antarctic ice streams (0.65 km/y) (Pfeffer et al., 2008). If ice discharge into the oceans from this region continues to increase, the sea-level rise contribution from western Antarctica will be significant. Pfeffer et al. (2008) estimated the potential sea-level rise contribution from Antarctica by 2100 by (i) doubling the velocities of Pine Island and Thwaites Glaciers within the first decade, (ii) accelerating the rate of ice loss by surface mass balance from the Antarctic Peninsula at the present-day rate of surface mass balance change, and (iii) scaling dynamic losses from the Antarctic Peninsula to losses by surface mass balance with a ratio of 1.31 (as observed in Greenland). The resulting projected sea-level rise is 0.15 m by 2100 (this case is also referred to as the 'Low 1' scenario).

Under the 'Low 1' scenario, the Greenland and Antarctic ice sheets would together contribute up to 0.31 m to mean sea-level by 2100 (Pfeffer et al., 2008). Under a more extreme scenario, 'High 1', that envisages more dramatic increases in dynamic losses, contributions to sea-level rise by 2100 of 0.54 m from Greenland (0.47 m from ice dynamics alone) and 0.62 m from Antarctica (all but 0.01 m from ice dynamics) are projected (Pfeffer et al., 2008), a total of nearly 1.2 m.

According to Radić and Hock (2011), the projected multi-model mean sea-level rise contribution from all mountain glaciers and ice caps on Earth is 0.12 ± 0.04 m by 2100. This is larger than an estimated contribution from surface mass balance of 0.08 m (Pfeffer et al., 2008). In the Radić and Hock study, mass loss by calving is neglected. Pfeffer et al. (2008) suggested that ice discharge from fast-flowing tidewater glaciers and ice cap outlets could add between 0.09 m (Scenario 'Low 1') and 0.47 m (Scenario 'High 1') to global mean sea level by 2100. Adding in the steric contribution to projected sea-level rise estimated by the IPCC AR4 Working Group 1 (Meehl et al., 2007) of 0.30 ± 0.15 m by 2100, the sum of land ice contributions to global mean sea level by 2100 would range between 0.79 m (Scenario 'Low 1') and 2.01 m (Scenario 'High 1') (Pfeffer et al., 2008). It should be emphasized, however, that these estimates are not the result of rigorous projections, but of scenarios designed to bracket the potential range of sea-level rise that could occur as a result of changes in ice dynamics.

As an alternative method for projecting the combined contributions of surface mass balance and ice dynamics to 21st century sea-level change, Rahmstorf (2007) and Grinsted et al., (2009) made semi-empirical projections by linking past sea level changes to observed surface air temperatures (see Figure 11.6). Their statistical projections of sea-level rise for the range of temperature projections associated with the IPCC A1B emissions scenario (2.3 to 4.3 °C temperature increase for 2100) range from 0.97 to 1.56 m above 1990 sea level by 2100. This is consistent with what happened during warming induced by variations in Earth's orbit during the last interglacial period (Overpeck et al., 2006).

Total projected sea-level rise resulting from all sources by 2100 cannot be estimated with high confidence at present. Lower and upper limits have a range of 0.79 to 2.01 m. A range of 0.90 to 1.6 m is considered the more plausible current estimate by the authors of the SWIPA assessment.

Finally, it should be emphasized that sea-level changes are far from equal from region to region as changes in the Earth's rotation and gravity field, ice load, ocean temperatures, ocean circulation, and fresh water supplies also impact the local mean sea level (Cazenave et al., 2008; Cazenave and Llovel, 2010). Between 1993 and 2007 the trend in observed sea level varied spatially from -5 mm/y to 15 mm/y. Taking into account the influences of changes in the Earth's rotation, the gravity field and related effects for the period 2000 to 2008, Bamber and Riva (2010) suggested that mass loss from Greenland resulted in local sea-level rise around South America, and local sea-level fall in much of eastern Canada and northwestern

Europe. Trends in local mean sea level will also depend on location relative to regions where ice mass loss is occurring due to the isostatic response (Milne et al., 2009). Overall, there is a tendency toward falling local sea level in the vicinity of regions where large decreases in ice mass are occurring.

11.2.6. Impacts of sea-level rise

The impacts of sea-level rise will be felt through an increase in mean sea level as well as through more frequent extreme sea-level events, such as storm surges. Impacts include increased severity and frequency of flooding in low-lying areas, salt water intrusion into surface water bodies and fresh water aquifers, beach erosion, and damage to infrastructure, agriculture, and the environment, including wetlands, intertidal zones, and mangroves, with significant impacts on biodiversity and ecosystem function. Hundreds of millions of people live in low-lying coastal areas, including heavily populated deltas such as Ganges-Brahmaputra (Bangladesh), the Rhine (Netherlands), and the Mekong (Vietnam) (Nicholls and Cazenave, 2010).

Sea-level rise severely threatens the existence of a number of small island states. They are in danger of becoming completely submerged or facing increased coastal flooding. Several of these states, such as the Maldives in the Indian Ocean, the Bahamas in the Caribbean, and Tuvalu in the Pacific, are topographically flat and at or near sea level. Others are volcanic with only narrow coastal strips that are habitable. Sea-level rise (especially if coupled with storm surges from stronger and longer lasting tropical storms) would have a great impact on the socio-economic condition of small islands and small island states. Adaptation will be challenging in many island settings, and in some places may not be feasible at all.

Changes in ocean circulation due to ocean temperature rise and freshening of sea water may affect regional sea levels, particularly in the northeastern United States, eastern China, and Japan, where sea level is projected to rise more than the global average. This would place large coastal cities, such as Shanghai (China) and New York City (USA) at greater risk of permanent inundation and coastal flooding.

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Section 11.3 – Pending technical/linguistic editing

Section 11.4 – Pending technical/linguistic editing

11.5. Observational Needs and Knowledge Gaps for the Cryosphere

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Key findings

Cryospheric observations help understand, assess, predict, mitigate, and adapt to climate variability and change; improve weather forecasting; help reduce the loss of life and property from disasters; provide a better understanding of environmental factors affecting health; and improve the management of resources. Closing gaps in snow and ice observation networks and in knowledge of cryospheric processes would allow the full potential of snow and ice information to be realized for the benefit of society. This assessment has made the following findings:

- Many Arctic surface-based observation networks for snow, lake and river ice, permafrost, and precipitation have diminished or been completely lost. Many important observations and monitoring efforts are research based and in need of sustained funding.

- There is no pan-Arctic dataset of in situ snow measurements. There are few measurements of snow depth on sea ice. Space-based capabilities for snow extent are robust, but methods of estimating snow water equivalent and snow depth are limited.

- Precipitation gauge networks are the most important source of information on high latitude snowfall but have large errors. Satellite methods are challenging but promising.

- While satellites have provided reliable observations of sea-ice extent, concentration, and motion for over 30 years, methods for estimating ice thickness from space are only now being developed. In situ measurements of ice thickness are sparse.

- There are numerous boreholes that provide temperatures for permafrost studies but many of the records are discontinuous and short. There is a wealth of historical data extending back 50 to 100 or more years; data rescue efforts are needed.

- Glacier inventories have basic data for less than half of the world's glaciers. Satellite-based inventories are discontinuous. Mass balance measurements over long periods are available for only a small subset of Arctic glaciers.

- Since the Arctic Climate Impact Assessment (ACIA) there has been considerable progress in the ability to estimate the mass balance of ice sheets. Gravity, laser altimetry, and radar data from satellites have become a major source of information. These new observational methods have broadened the range of estimates, revealing that uncertainties may be larger than previously thought.

- Traditional knowledge can provide important value-added content to data products and serves to make the data more relevant to northern users. Community-based observing programs provide a mechanism for two-way knowledge transfers.

- There are considerable uncertainties in modeling cryospheric processes. Permafrost models under-represent ice content and the insulating effect of the organic layer; climate models do not resolve the steep topography of the Greenland Ice Sheet margins; models of snow-vegetation interactions need to be improved; and models that link meteorology to glacier mass balance need to incorporate downscaling techniques and satellite data.

11.5.1. Introduction

Snow and ice observations play a crucial role in numerical weather prediction, climate monitoring, change detection, hazard forecasting, and water resource management. A comprehensive set of observations is required to assess the efficacy of mitigation and adaptive strategies and to prioritize needs for governance. Improved mapping and monitoring of drifting ice and icebergs, local sea-ice formation, annual variations in ice cover, and trends in storm surge characteristics are important to shipping, industrial developments, and

local hunting. Forecasts of break-up and freeze-up of the ice will help industry to determine the seasonal window for operations. Monitoring of snow cover, winter thaw, and rain will provide information relevant for hydropower, drought, snow load, avalanches, and floods. For many hunting activities, knowledge of changes in the timing and amount of snow and ice are important. For maintaining the integrity of infrastructure, monitoring of permafrost characteristics is important.

However, conditions in areas where the cryosphere exists are harsh, and in situ observations there are difficult and expensive. Satellite monitoring overcomes some of the logistical obstacles, but satellites are costly and do not yet fully address the range of geophysical variables needed to understand the cryosphere. Furthermore, satellite products will always require ground truth for validation. A comprehensive cryosphere observing system must therefore be a combination of ground-based instrumentation, satellite remote sensing, aircraft measurements, modeling, and data management (Figure 11.5.1). Surface and airborne observations provide data that cannot currently be measured from space, more detailed information in critical areas, and observations with which to calibrate and validate satellite retrievals. Satellite instruments are essential for delivering sustained, consistent observations of the global Arctic cryosphere and are key to extending local in situ measurements. No single all-encompassing sensor exists; rather, the combination and synthesis of data from different yet complementary sensors is essential and underlines the critical importance of maintaining key synergetic elements of the system.

While the previous chapters use snow and ice data to assess changes in the Arctic cryosphere, Section 11.5 assesses the state of the cryosphere observing system itself. It addresses observations more broadly, synthesizing the material presented earlier and focusing on observational gaps. Table 11.5.1 provides a summary of the observing systems for snow and ice. It lists the measurement approach for the major variables, describes the status of the networks (operational or research), provides a qualitative assessment of how well each is meeting the measurement requirements, and outlines any major issues. As robust as the satellite and in situ measurement networks are, there are many shortcomings in the cryosphere observing system that give rise to sometimes-large uncertainties in assessments such as SWIPA. These are discussed for each cryospheric element in the following sections. The societal impacts of changes in the cryosphere, and hence the importance of robust measurements of snow and ice properties, are described in the socio-economic synthesis chapter.

Table 11.5.1. Observational readiness of snow and ice measurements for the observing system overall. Green satisfies the requirements (≥ 85%); cyan meets the requirements most of the time (≥ 70%); yellow meets the requirements some of the time, or only for specific conditions; red does not meet the requirements. O: operational, R: research, C: commercial, L: long-term (≥ 20 years) record.

	In situ	**Satellite**	**Major gaps in observations**
Ocean			
Ice extent	coastal radar (R), ship observations	passive microwave (O, L)	In situ coverage is sparse and incomplete
Ice concentration	ship observations	passive microwave (O, L)	Potentially large uncertainties in satellite retrievals in summer
Ice thickness	ice-profiling sonar on moorings (O); mass balance buoys (C), electromagnetic sleds (R)	optical, laser and radar altimeter (R)	Satellite methods are still developing; snow depth on ice is an unknown
Ice motion	drifting buoys (O, L); coastal radar (R)	passive and active microwave (O, L); optical (R)	Important small-scale motions not captured by satellites; in situ measurements sparse
Snow depth on ice	depth gauge (R)	passive microwave (R)	Satellite method is limited to first-year ice with potentially large uncertainties; in situ data are sparse
Sea level	tide gauges (O, L); bottom pressure recorders (C)	altimeters (R)	

Surface temperature	drifting buoys (O, L)	optical (O, L)	Uncertainty in satellite estimates due to cloud cover
Albedo	radiometers (O, L)	optical (O, L)	Sparse in situ coverage; significant uncertainty
Terrestrial			
Snow cover	depth gauge (O, L)	optical (O, L)	In situ network is declining
Snow depth	depth gauge (O, L)	optical (R)	Satellite method is limited to tall-grass prairie
Snow water equivalent	various methods (O)	passive microwave (R)	In situ coverage is sparse
Freshwater ice	visual observations (O, L)	optical (R)	Declining observation network
Glacier, ice cap, ice sheet thickness	seismology (R)	radar (R), gravity (R)	Sporadic coverage
Glacier length, area	surveys (R, L)	optical (R)	Incomplete coverage
Glacier, ice cap, ice sheet motion	GPS (R)	InSAR, visible imager (R)	Sporadic coverage
Permafrost: ground temperature	boreholes (O, L)	(none)	Large portions of the Arctic not covered
Permafrost active layer thickness	boreholes, probes (O, L)	(none)	Large portions of the Arctic not covered
Surface temperature	thermisters, thermocouples (O, L)	optical (O, L)	Satellite method is clear sky only
Albedo	radiometers (O, L)	optical (O, L)	Sporadic in situ coverage; significant uncertainty

11.5.2. Sea ice

The first quantifiable observations of sea-ice properties came from ships sailing in and near ice-covered regions as early as the 1500s (Figure 11.5.2). National ice services began routinely producing ice charts for the Arctic in the 1950s using visual aerial reconnaissance missions and aerial photography. The thickness of ice fragments laid on edge due to the icebreaker movement has been determined using onboard digital video cameras on Russian icebreakers for many years. Submarine sonar observations of ice draft under the Arctic sea ice began in earnest in the 1950s, although the data are sparse and remained classified for decades. Moored upward-looking sonar (ULS) gives thickness and volume flux estimates from below. Electromagnetic (EM) sensors mounted on ships, surface vehicles, helicopters, or aircraft can estimate thickness. Surveys using new airborne and satellite lidar methods can now provide thickness estimates for large regions, although uncertainties, particularly in snow cover, limit accuracy. More recently, near-surface observations have been taken from autonomous underwater vehicles and from unmanned aerial vehicles, although their small size limits the type of sensor that can be carried. Manned Russian 'North Pole drifting stations' have operated on the Arctic sea ice from 1937 to 1991 and since 2003. Mass balance buoys are now being deployed to obtain ice and snow thickness and internal temperatures.

Satellite-borne visible and infrared sensors provide observations of surface albedo and temperature, snow cover, radiative fluxes, ice extent, concentration, motion, and melt, but coverage is limited due to frequent cloud cover, and for visible imagery, sunlight. Satellite passive microwave radiometers provide a 30-year plus record, with near-complete daily coverage, of the polar regions under all sky conditions for concentration and extent, motion, and melt. Unfortunately, the spatial resolution of these products (6 to 25 km) makes it impossible to obtain detailed information on the ice cover, such as deformation, melt-pond and lead formation, and ridging. Synthetic aperture radar (SAR) instruments provide high-resolution information on deformation, leads, ridging, and new ice production, which is important for marine transportation.

Scatterometry provides information at a similar spatial scale to passive microwave data, but can provide better information on perennial ice cover and complementary information on other properties (e.g., melt and snow). Spaceborne laser and radar altimeters can provide indirect estimates of ice thickness, which is useful for navigation and for modeling ice-atmosphere fluxes. Accurate estimates of ice thickness from altimeters require good knowledge of snow depth and density. More in situ measurements of snow depth are needed for

better estimates of ice thickness (Sea Ice Thickness chapter). The potential to derive snow depth on floating ice by passive microwave remote sensing has been demonstrated by Markus and Cavalieri (2000), but it is only applicable over first-year ice regimes and requires further research. Other satellite-based products for estimating ice thickness and age are currently being developed.

There is no single accepted passive microwave ice concentration product[1], leading to confusion among users. Rigorous evaluation and consolidation of products are needed, together with formal estimates of algorithm uncertainties and errors. This should include improved inter-sensor calibration with longer periods of overlap, rigorous evaluation and intercomparison of current algorithms, and the development of data fusion methods to obtain optimal combined products. Sea-ice observations are not well suited for model assimilation because their error structures are not well known at the grid cell or pixel level.

11.5.3. The Greenland Ice Sheet

The Greenland Ice Sheet has been studied extensively in an effort to quantify the surface mass balance and to estimate its contribution to global sea level. However, in situ observations are limited due to the large area and remoteness of the ice sheet. The notable exception is the record from the Greenland Climate Network (GC-Net) of automatic weather stations, operating since 1995 and distributed across the ice sheet. Scientists must therefore rely on remotely sensed observations and estimations from downscaled general circulation models (GCMs) or high-resolution regional climate models (RCMs) to investigate surface mass balance. However, systematic biases in the models relative to the in situ record illustrate that the climate of the ice sheet is more or less decoupled from the coastal regions where most of the direct observations that feed the models are made.

Satellites have triggered a quantum leap in observations of the ice sheets (Chapter 8, this volume). Images acquired by reconnaissance satellites starting in 1962 provide a treasure of historical data for gauging subsequent changes in the positions of shear margins, grounding lines and glacier terminuses (Kim et al., 2001; Zhou and Jezek, 2002). Passive and active microwave observations are used to measure the onset, duration, and extent of ice sheet surface melt. Passive microwave data have also been used to estimate average annual accumulation (Zwally and Giovinetto, 1995). Satellite altimeters are invaluable for measuring ice sheet elevation and also provide information on subglacial water movement (Wingham et al., 2006; Fricker et al., 2007). SAR interferometry (InSAR) provides a powerful means to investigate glacier flow. Satellite gravity instruments provide measurements of ice sheet mass change (Chen et al., 2006; Velicogna, 2009).

In situ measurements are progressing, although at a slow pace due to resource-demanding fieldwork (Chapter 8, this volume). Snow pits are used to measure near-surface temperature, snow and firn density, crystalline structure, surface accumulation, and when coupled with GPS measurements, surface velocity. In situ seismological experiments are used to measure ice thickness, near-surface density, and the properties of the glacier bed. Surface and airborne radars are used to measure ice thickness, although spatial coverage is limited and there are still large gaps in the ice thickness maps. To efficiently fill these gaps, fixed-wing aircraft, unmanned aerial vehicles, and satellite implementations of ice sounding synthetic aperture radars should be explored. A combined effort in advancing process theory, field methods, coverage, and modeling efforts could facilitate major advances in understanding ice sheet mass balance and help create less uncertain future scenarios of ice mass loss and sea-level rise.

Ice sheet models are insensitive to ocean forcing, in contrast to recent observations of significant velocity variations in marine-based outlet glaciers, which were most likely triggered by ice-ocean interactions and thinning at the grounding line. Gaps in knowledge are mainly related to physical processes of fast-flowing outlet glaciers and ice streams, bed topography, and melting rates under floating glacier tongues. Melt rates on the submarine parts of terminal ice cliffs are also poorly known and often ignored.

Compared with processes that take place at the base of the ice sheet, those that influence the surface mass balance are relatively well understood and considerably easier to observe. Despite this, reconstructions of the individual components that make up the surface mass balance differ significantly. The differences can be as large as the inferred increase in mass loss due to changes in ice dynamics over the past decade. The

[1] The term 'product' as used in Section 11.5 does not imply an extensive level of validation nor the adherence to any particular standard. The products may be research or operational datasets, but all have undergone at least a basic level of validation.

uncertainties in the surface mass balance are largely a result of the paucity of spatially extensive in situ observations. Reducing uncertainties will require a concerted effort to collect more targeted in situ data, especially from the percolation and ablation zone. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow, and subgrid-scale effects will also be needed. Meteorological measurements from automatic weather station networks are the key means of estimating near-surface fluxes. These measurements should be maintained and spatially extended to key regions on the Greenland Ice Sheet.

Instead of reducing uncertainty about changes in the mass balance of the Greenland Ice Sheet, new observational methods from space have broadened the range of estimates, revealing that the uncertainties of methods may be larger than previously thought. They also point to the incomplete understanding of the fundamental dynamics of calving glaciers and land-terminating outlets of large ice sheets. Most changes appear in the marginal regions of the ice sheet where data are scarce. Many climate models do not resolve the steep topography of the Greenland Ice Sheet margins, limiting skill in simulating orographic precipitation and ice sheet ablation. New progress in snow physics theory and parameterizations needs to be incorporated into climate models (Chapter 8, this volume).

Climate models generally do not include ice sheets and glaciers, a lack that limits their use for projecting sea-level rise. Particular challenges in modeling sea-level rise are the coupling of ice sheet and ice bed and of ice sheet, ice shelf, and ocean. Key uncertainties in predicting Greenland's contribution to sea-level rise are ice dynamics and surface mass balance. The response to warming remains a large uncertainty.

11.5.4. Terrestrial snow

Surface-based methods of measuring snow properties include graduated probes, acoustic depth sounders, coring devices, snow pillows (pressure transducers), and excavation of snow pits with detailed sampling. Worldwide, many surface-based snow-observation networks have diminished or have been completely lost. The number of Arctic stations reporting snow depth increased between 1960 and the 1990s, but decreased in many areas over the past decade. The most common observation is snow depth. Far fewer snow-water equivalent (SWE) observations are available, and observations of other snow parameters (e.g., layer snow temperature, layer boundaries, hardness, grain form, grain size, density, strength, and stability) are limited. A coordinated plan for surface-based snow observation networks must be developed, first at the national, then at the international level. Although there are regional snow cover products (Chapter 4, this volume), at present there is no pan-Arctic dataset of in situ snow measurements, and there is no single archive that has all snow depth data. Data are particularly sparse for the mountain and tundra regions.

Observations of snow extent from visible, near-infrared, and microwave satellite sensors have been widely used since 1966. Space-based capabilities for observing snow depth and SWE are more limited. Microwave sensors appear ideal for this purpose, but current sensors lack optimal combinations of frequencies and spatial resolutions, particularly for mountainous terrain. Long-term SWE data have been derived from passive microwave sensors, although comparative analyses of basin-wide SWE and winter precipitation suggest uncertainties in the determination of snowfall amounts and SWE over large watersheds and regions with different physical characteristics (Yang et al., 2009). SAR observations overcome the resolution problem, but current sensors operate at frequencies too low to be useful for most snowpacks. Priority should be given to research and the development of algorithms and new sensors to measure SWE (high frequency SAR), under a wide range of vegetation conditions.

Energy and mass-balance snow models, now widely used for many applications, are highly nonlinear and include several interacting variables. These models are under-constrained by mass-oriented observations of SWE or snow depth alone. Snow temperature observations are necessary to constrain the energy states of these models. Snow observation sites are often not representative of surrounding areas. This results in observation biases that can be easily propagated to large regions through modeling and data assimilation. Such biases are common and can be quite large.

Further work is needed to improve models of snow-vegetation interactions. There are dramatic differences in snow model response if vegetation is included. One of the problems is that parameterization of key processes such as snow unloading are either non-existent or based on site-specific data. Integrated multi-sensor data fusion and global analysis systems that blend snow observations from all sources must be improved. The ideal global snow observing system will use observations from all relevant sources in coherent, consistent

high-resolution analyses of snow extent, depth, SWE, wetness, and albedo. No current system provides global coverage.

11.5.5. Permafrost and seasonally frozen ground

The Global Terrestrial Network for Permafrost (GTN-P) identifies permafrost thermal state (i.e., ground temperature) and the active layer thickness as key variables for monitoring (WMO, 1997). Many permafrost temperature records are of short duration and discontinuous, although some sites have 20- to 40-year time series. Currently about 400 boreholes are listed in the GTN-P database; around half of these are in Arctic countries (Figure 11.5.3). There are currently over 100 Arctic sites that report seasonal thaw, soil temperatures, moisture and frost heave, and subsidence.

No formal international network exists for seasonally frozen soil temperature or moisture observations, although several programs make some data available. The Arctic Coastal Dynamics (ACD) program began in 1999, where a network of 25 key sites was identified along the coastline of the entire Arctic Basin (Rachold et al., 2005). Some national networks are extensive. There were more than 800 stations in the former Soviet Union where soil temperature and soil freezing depth were measured on a daily basis (Chudinova et al., 2006), but the number of stations has declined significantly in the past two decades. Canada and the United States have many stations where soil temperature is measured, although not all are Arctic.

Many of the GTN-P sites are not maintained for long-term monitoring. There are significant thematic and regional gaps in the present networks, especially in eastern and central Canada, most ice-free areas in Greenland, and north-central and northeastern Russia. Many sites lack instrumentation for collecting the ancillary data required to analyze the linkages between permafrost and climate. Existing borehole and active layer networks must be expanded. Soil temperature and frost depth measurements should be standard parameters for all cold-region meteorological stations.

Properties of permafrost terrain are currently not directly detected from remote sensing platforms. Zhang et al. (2004) and Duguay et al. (2005) provided a comprehensive overview of satellite remote sensing of permafrost and seasonally frozen ground. Surface indicators of permafrost terrains that are discernable by remote sensing include pingos, thaw lakes and basins, thaw slumps, thermo-erosional valleys, thermokarst mounds, ice wedge polygons, beaded drainage, slope failures, and rock glaciers. SAR data have been used to map zones of wintertime heat loss that indirectly approximate the distribution of permafrost (Granberg, 1994). Scatterometer data have been used to monitor freeze and thaw cycles (Bartsch et al., 2007). Data from passive microwave sensors can be used to detect surface-soil freeze or thaw status (Zhang and Armstrong, 2001) as well as soil moisture. The application of multi-temporal, basin-scale gravity data for the detection of mass loss from ground ice melting in lowland permafrost regions should be evaluated.

There is a wealth of historical observations extending back in time for 50 to 100 or more years. For example, the former Soviet Union has some sites with daily observations extending back to the 19th century. However, many of the historical national databases lack adequate metadata. While data rescue is labor intensive and can be expensive, it would provide a valuable historical perspective, putting cryosphere-climate interactions into the current context of rapid change (Chapter 5, this volume).

There are considerable uncertainties in modeling future permafrost distribution and dynamics. These include an under-representation of the ice content and the organic layer and its importance in insulating permafrost during climatic warming. Permafrost models also fail to adequately represent the disequilibrium that has arisen because some current permafrost is related to past climates. This results in a lag period between a climatic change and a response of the permafrost.

11.5.6. Glaciers and ice caps

For glaciers and ice caps, key uncertainties are the incomplete knowledge of their total mass and the regional variations in mass balance. New datasets on ice thickness and glacier outlines are required to improve estimates of glacier melt. The World Glacier Monitoring Service (WGMS) combines in situ measurements of mass balance and length change with global information from remote sensing data (Haeberli, 2006). Systematic compilation of inventories at regional or national scales started in the 1940s based on aerial photography. Of approximately 160 000 glaciers, about 44% (71 000) are currently stored in the World Glacier Inventory (WGI) (WGMS, 1989). Many regions are still not included in the WGI although data have been compiled.

High-resolution multispectral optical sensors are the most efficient means for glacier mapping. Repeat inventories are needed at five- to ten-year intervals for global change studies and assessing change of water resources. The Global Land Ice Measurements from Space (GLIMS) Project (Bishop et al., 2004; Kargel et al., 2005) will establish a digital baseline inventory of ice extent during the period 2000 to 2005 for comparison with inventories created at earlier and later times. This will in turn lead to improved information on the global extent of glacier retreats and advances. Enhanced resolution data from satellite scatterometers are used to generate products such as melt onset and freeze-up dates, melt duration, and inferred facies distributions for ice caps and large glaciers. InSAR can be used to map glacier surface topography and provide maps of surface motion. Speckle tracking with visible imagery is also important for motion estimation and is more widespread than InSAR approaches, as interpretation of interferometric velocities is not straightforward. The use of satellite gravity measurements for estimates of mass change for large glaciated regions should be expanded.

Continuous annual mass balance measurements made over relatively long periods from field measurements of accumulation and ablation are only available for about 50 glaciers worldwide (Dyurgerov and Meier, 1997). Many glacier regions have no mass balance observations, which biases the network of mass balance sites strongly toward those glaciers that are easily accessible, particularly smaller glaciers in Europe, Scandinavia, and North America. There are issues with measurement methods being used: such as whether data for winter and summer balances are available or only the annual balance, whether processes of internal accumulation are taken into account in the calculation of annual balance, and whether changes in glacier area over the period of record are taken into account in estimating the annual glacier-wide balance (Chapter 7, this volume).

For the most part, ice thickness measurements come from surface-based radars, as borehole studies of ice deformation properties and of subglacial till are rare. However, some of the ice masses of the Arctic are too large to survey with surface radars, so airborne surveys are important. Ice coring in the cold firn of glaciers and ice caps is needed to extend the climate archive obtained by coring on the ice sheets (Chapter 7, this volume).

The glacier topography database is fragmentary and of poor quality. Space-based data are needed to improve it. A dedicated InSAR mission for precise mapping of glacier topography is a high priority for determining the evolution of changes. A satellite mission providing information on accumulation should be implemented. Repeat laser altimetry surveying should ideally be extended to all major glaciated regions of the Arctic.

Models that link meteorology to glacier mass balance and dynamic response need to be improved. Downscaling techniques need to be developed for feeding such models with GCM data. Remote sensing data are needed to initialize and validate these models. Water management tools for glacier runoff will also benefit from these developments.

11.5.7. Freshwater ice

Freeze-up and break-up dates from lakes and rivers have traditionally been determined from surface observations. Unfortunately, surface-based lake and river ice observations have been steadily declining since the 1980s to a point where networks have almost disappeared in many countries (Figure 11.5.4). In Canada, freeze-up and break-up observations were reported for only 12 water bodies across the entire country during the 2000/01 season. Some lake and river ice observations from various countries have been compiled into the Global Lake and River Ice Phenology Database (GLRID). The database contains records for 748 sites, but the number of sites reporting ice observations has plummeted since the 1980s. Financial cutbacks and the automation of meteorological stations in the vicinity of lake and river ice observation sites are the two main reasons for the drastic decline in the surface-based networks. A set of target regions and lakes and rivers must be identified for future long-term ice monitoring. A major data rescue effort must be undertaken (Chapter 6, this volume).

The Canadian Ice Service (CIS) began operational, weekly monitoring of ice extent on large lakes in 1995 using visible/IR and SAR satellite data. The program started with 34 lakes in 1995 and now has 136 lakes, mostly in Canada with a few in the United States. It is possible to derive dates of complete freeze-over and when water is clear of ice with an accuracy of ± 1 week using this dataset. In 1997, NOAA began generating a daily snow and ice product with the Interactive Multisensor Snow and Ice Mapping System (IMS). The IMS incorporates a wide variety of satellite imagery as well as derived mapped products and surface observations.

The coarse resolution of the 24-km product allowed for mapping of ice extent only on the largest lakes of the Northern Hemisphere, but a 4-km resolution product became available starting in 2004.

Lake-ice thickness measurements have primarily been through field observation programs. Like the freeze-up and break-up network, ice thickness networks are sparse, and the number of sites has been steadily declining since the 1980s (Lenormand et al., 2002). Ice thickness has been estimated with some success using numerical ice growth models (see Duguay et al., 2003) and through the synergistic use of optical and SAR data on shallow Arctic lakes (Duguay and Lafleur, 2003). Passive microwave data offer a promising means to obtain lake-ice thickness from very large lakes. It might also be possible to obtain the total thickness of snow and ice on medium to large lakes from snow surface elevation data acquired by spaceborne lasers (Jeffries et al., 2005).

Knowing which lakes and rivers freeze to their bottom (grounded ice) or not (floating ice), and when this occurs during the course of the winter period is critical for mapping and monitoring water availability and fish over-wintering habitats and for planning winter ice roads. The potential of SAR imagery has been demonstrated for such purposes (Jeffries et al., 2005; Duguay and Lafleur, 2003). Surface measurements of parameters such as the dates of the first appearance of ice, complete freeze-over, beginning of thaw, and when the water body becomes completely free of ice (which are important for transportation) are not currently available.

11.5.8. Solid precipitation

Methods of observing solid precipitation include the precipitation gauge network, satellite remote sensing, and ground radar. Precipitation gauge networks and data are long term and fundamental; they make it possible to define global snowfall and climate regimes as well as changes. Manual and automatic precipitation gauges can measure the water equivalent of snowfall but not snow particle size. Manual gauges are standard at most national networks, although the use of automated systems is growing rapidly. Snow rulers are also used for snowfall observations in national or regional networks. They provide snow depth information only, not SWE. As important as gauge measurements are, quality control can be difficult (Figure 11.5.5).

Estimating solid precipitation amounts and rates from space is a challenge. The use of satellite data for solid precipitation measurements is limited and is an active area of research. Experimental products are available from passive microwave sounders (Kongoli et al., 2003) and cloud radar (Hiley et al., 2010). Various satellite data have been merged to improve the spatial and temporal coverage of precipitation products (Huffman et al., 1997). The future precipitation radar of the Global Precipitation Mission will observe solid precipitation, although its utility for high latitude snowfall is not clear. Of course, remote sensing of snow-cover extent and SWE data provide information on snowfall and have been widely used for large-scale climate and hydrology analyses.

Satellite retrievals have been blended with gauge data. For instance, the Global Precipitation Climatology Project has provided global coverage since 1987. Examination of hydrological budgets over several global river basins revealed unrealistically low precipitation in the merged products (Fekete et al., 2004). Over land, the magnitude of these combined products is primarily dominated by gauge observations. The under-estimation of land precipitation in the gauge-based analyses, and thus the merged analyses, is mostly due to the combined effect of a sparse gauge network biased toward low elevations, a lack of consideration of orographic effects, and gauge undercatch of precipitation, especially snowfall in cold and windy conditions.

Although ground-based radar provides information about precipitation at high spatial and temporal resolution, the coverage it provides is limited – much less extensive than the gauge networks. The radar network is particularly sparse and expensive to operate at high latitudes. In complex terrain, mountains often block the radar beam or the radar is located to scan over the top of mountains and not in the valleys. A new innovation is the deployment of a network of redundant low-cost, low-maintenance radars to scan the low levels of the atmosphere.

Problems in current precipitation gauge networks include the sparseness and decline of precipitation observation networks in cold regions, uneven distribution of sites and a resulting bias toward coastal and low-elevation areas, spatial and temporal discontinuities of precipitation measurements induced by changes in observation methods (such as automation) and by different observation techniques used across national borders, biases in gauge measurements (such as wind-induced undercatch), underestimation of trace and low

amounts of precipitation, and blowing snow in the gauges. There is no operational precipitation network over the oceans.

11.5.9. Summary

While there are numerous snow and ice surface measurement sites across the Arctic, operations at existing stations are, in general, not well coordinated. There is a need to improve the coordination of resources provided by national and international agencies responsible for cryospheric observations, and to facilitate the transition of research-based products into sustained monitoring systems. There is also a need to standardize the types and methods of measurements at surface stations, so that a robust and consistent set of snow and ice properties is available across the Arctic.

The satellite observing system for the cryosphere is robust, and missions planned for the next 10 to 20 years will provide even greater capabilities. Figure 11.5.6 shows the timeline of satellites that are used for monitoring the cryosphere. However, some potential gaps will be detrimental to long-term monitoring. In particular, the current gap in laser altimetry and the potential near-future gap in gravity measurements will impact ice sheet and glacier monitoring and change assessment. Additionally, there are some critical parameters that are difficult to measure from space, notably sea-ice thickness, snow water equivalent, and accumulation on glaciers and ice sheets.

Traditional knowledge can provide important value-added content to data products and information and serves to make the data more useful and relevant to northern users (Chapter 10, this volume). Community-based observing programs provide a mechanism for two-way knowledge transfer, such as the Gwich'in First Nation International Polar Year (IPY) environmental monitoring project that includes snow cover monitoring along with a wide range of environmental, wildlife, and cultural variables. The ELOKA (Exchange for Local Observations and Knowledge of the Arctic) project will facilitate the collection, preservation, exchange, and use of local observations and knowledge of the Arctic and will foster collaboration between local and international researchers. Similarly, Sámi and scientists in Scandinavia are uniting traditional knowledge and scientific information on snow properties and monitoring (Chapter 4, this volume). Traditional knowledge can provide important supplementary information for the assessment of changes in the Arctic cryosphere.

While there are gaps and deficiencies in the cryosphere observing system, the data that do exist are underutilized. The use of cryospheric products in weather and climate models needs to be extended. Surface albedo is a good example, as many models use rudimentary and unrealistic albedo schemes for snow and ice. Data assimilation schemes for snow and ice variables are generally underdeveloped. Sea-ice extent and concentration data are assimilated at several real-time modeling centers, but ice thickness and motion data are not. Reprocessing of data products should be organized and promoted by national and international agencies and organizations.

Acknowledgements. Co-authors from other sections of this report provided valuable input, notably Lars-Otto Reiersen, Margareta Johansson, Terry Callaghan, Walt Meier, and Jim Reist. Some of the material in this chapter is based on the IGOS Cryosphere Theme Report. We thank the authors of that report.

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