Snow, Water, Ice and Permafrost in the Arctic (SWIPA):

Climate Change and the Cryosphere

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Production

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Acknowledgments

Preface

This report presents the findings of the *Snow, Water, Ice and Permafrost in the Arctic (SWIPA): Climate Change and the Cryosphere* assessment performed by the Arctic Monitoring and Assessment Programme (AMAP) in close cooperation with the International Arctic Science Committee (IASC), the World Climate Research Programme/Climate and Cryosphere (WCRP/CliC) Project and the International Arctic Social Sciences Association (IASSA).

The SWIPA assessment is the third AMAP assessment addressing Arctic climate issues and is a direct follow-up to the Arctic Climate Impact Assessment (ACIA)[1] published in 2005. The ACIA reviewed the state of knowledge regarding ongoing change in the Arctic climate and documented the role that the Arctic plays in the global climate system; it represents the benchmark against which this updated assessment of change in the Arctic cryosphere has been developed.

The SWIPA assessment was conducted between 2008 and 2011 by an international group of over 200 scientists, experts and knowledgeable members of the Arctic indigenous communities. Lead authors were selected by an open nomination process coordinated by AMAP, IASC, WCRP/CliC and several national and international organizations. A similar process was used to select international experts who independently reviewed this report. A SWIPA Integration Team, including the lead authors for the twelve chapters, was responsible for scientific oversight and coordination of all work related to the preparation of the assessment report. Documentation available on the website www.amap.no includes listings of the comments received from the peer reviewers.

Information contained in this report is fully referenced and based first and foremost on research and monitoring efforts published since 2003 (i.e., information gathered since the ACIA report was undertaken). It includes peer-reviewed material accepted for publication up until October 2010, and in some cases later. Unpublished monitoring information, including both *in situ* and satellite observations, with well-established national and international standards and QA/QC (quality assurance / quality control) protocol are also part of the assessment. Acknowledging national differences in scientific quality assurance, the SWIPA assessment therefore draws mainly on peer-reviewed publications and work accepted for publication in respected scientific journals, including works reviewed by Russian scientific committees. Other sources of information, such as government reports, design standards, official records, statistics and other publicly available material, have however also been included in the work in order to provide as complete a picture of the effects of a changing Arctic cryosphere as possible; this is particularly the case for parts of the assessment dealing with the human dimension. All such references have been collected and are available upon request (at cost of reproduction) from the AMAP Secretariat. Care has been taken to ensure that no critical probability statements are based on these materials.

Access to reliable and up-to-date information is essential for the development of science-based decision-making regarding ongoing changes in the Arctic and their global implications. SWIPA summary reports[2] and SWIPA films (available with different language subtitles) have therefore been specifically developed for policy-makers, summarizing the main findings of the SWIPA assessment. The SWIPA lead authors have confirmed that both this report and its summary report accurately and fully reflect their scientific assessment. This report constitutes the fully-referenced scientific basis for all statements made in the SWIPA summary report and its executive summary with recommendations for policy makers. The SWIPA reports and films are available from the AMAP Secretariat and on the AMAP website www.amap.no.

AMAP and its partner organizations would like to express their appreciation to all experts who have contributed their time and effort, and data; and especially to the lead authors and members of the SWIPA Integration Team who coordinated the production of this report. Thanks are also due to the many referees and reviewers who contributed to the SWIPA peer-review process and provided valuable comments that helped to ensure the quality of the report. A list of the main contributors is included at the start of each chapter. The list is not comprehensive. Specifically, it does not include the many national institutes, laboratories and organizations, and their staff, which have been involved in the various countries. Apologies, and no lesser thanks are given to any individuals unintentionally omitted from the list. Special thanks are due to the lead authors responsible for the preparation of the various chapters of this report.

The support of the Arctic countries and non-Arctic countries implementing research and monitoring in the Arctic is vital to the success of AMAP. The AMAP work is essentially based on ongoing activities within these countries, and the countries also provide the necessary support for most of the experts involved in the preparation of the AMAP assessments. In particular, AMAP would like to thank Canada, Denmark, Norway and the Nordic Council of Ministers for their financial support to the SWIPA work, and to sponsors of programmes and projects that have delivered data for use in this assessment. Special thanks are given to those experts involved in International Polar Year (IPY) projects who made their results available for the SWIPA assessment.

The AMAP Working Group is pleased to present its assessment to the Arctic Council and the international science community.

Morten Skovgaard Olsen (SWIPA Chair)
Russel Shearer (AMAP Chair)
Lars-Otto Reiersen (AMAP Executive Secretary)

Oslo, April 2011

Disclaimer: The views expressed in this peer-reviewed report are the responsibility of the authors of the report and do not necessarily reflect the views of the Arctic Council, its members or its observers.

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[1] ACIA, 2005. Arctic Climate Impact Assessment. Cambridge University Press.

[2] AMAP, 2012. Arctic Climate 2011: Snow, Water, Ice and Permafrost in the Arctic (SWIPA). Arctic Monitoring and Assessment Programme, Oslo, Norway.

Executive Summary and Key Messages

SWIPA Summary for policymakers

AMAP's new assessment of the impacts of climate change on Snow, Water, Ice and Permafrost in the Arctic (SWIPA) brings together the latest scientific knowledge about the changing state of each component of the Arctic 'cryosphere'. It examines how these changes will impact both the Arctic as a whole and people living within the Arctic and elsewhere in the world.

'Cryosphere' is the scientific term for that part of the Earth's surface that is seasonally or perennially frozen. It includes snow, frozen ground, ice on rivers and lakes, glaciers, ice caps, ice sheets and sea ice. The cryosphere structures the physical environment of the Arctic. It provides services to humans such as freshwater supplies and transport routes. The cryosphere is an integral part of the climate system, and affects climate regionally and globally.

The SWIPA Assessment follows on from the Arctic Climate Impact Assessment (ACIA), published in 2005. It aims to update the findings from ACIA and to provide more in-depth coverage of issues related to the Arctic cryosphere.

The observed changes in sea ice on the Arctic Ocean and in the mass of the Greenland Ice Sheet and Arctic ice caps and glaciers over the past ten years are dramatic and represent an obvious departure from the long-term patterns.

Some elements of the cryosphere, such as the extent of snow, ice over water, and the dynamics of glaciers and ice streams vary greatly over short timescales (seasonally, or from year to year) and from place to place. Other aspects of the cryosphere, such as the extent of permafrost and large ice sheets, vary and change over decadal time scales and large areas. Distinguishing long-term change from natural variability requires data to be collected at many locations over many years and carefully analyzed. Detecting these cryospheric responses to changing climate presents different challenges and requires long term records as well as high frequency observations.

Why the Arctic cryosphere is changing

The past six years (2005–2010) have been the warmest period ever recorded in the Arctic. Higher surface air temperatures are driving changes in the cryosphere.

Key finding 1

There is evidence that two components of the Arctic cryosphere – snow and sea ice – are interacting with the climate system to accelerate warming.

Key finding 2

The Arctic is warming. Surface air temperatures in the Arctic since 2005 have been higher than for any five-year period since measurements began around 1880. The increase in annual average temperature since 1980 has been twice as high over the Arctic as it has been over the rest of the world. Evidence from lake sediments, tree rings and ice cores indicates that Arctic summer temperatures have been higher in the past few decades than at any time in the past 2000 years. Previously unseen weather patterns and ocean currents have been observed, including higher inflows of warm water entering the Arctic Ocean from the Pacific. These changes are the main drivers of change in the Arctic cryosphere.

In attributing the cause of warming in the Arctic, SWIPA refers to the findings of the Fourth Assessment Report from the Intergovernmental Panel on Climate Change (IPCC). This states that "*Most of the observed increase in global average temperatures since the mid-20th century is very likely* [> 90% probability] *due to the observed increase in anthropogenic GHG [greenhouse gas] concentrations*".

Climate-cryosphere interactions may now be accelerating warming

The greatest increase in surface air temperature has happened in autumn, in regions where sea ice has disappeared by the end of summer. This suggests that the sea is absorbing more of the sun's energy during the summer because of the loss of ice cover. The extra energy is being released as heat in autumn, further warming the Arctic lower atmosphere. Over land, the number of days with snow cover has changed mostly in spring. Early snow melt is accelerated by earlier and stronger warming of land surfaces that are no longer snow-covered.

These processes are termed 'feedbacks'. Snow feedbacks are

well known. The sea ice feedback has been anticipated by climate scientists, but clear evidence for it has only been observed in the Arctic in the past five years.

A number of other potential feedback mechanisms at play in the Arctic have been identified. These mechanisms can alter the rate or even direction of climate change and associated changes in the cryosphere. Of those feedbacks expected to have strong effects, seven lead to further and/or accelerated warming, and just one leads to cooling. The intensity of feedbacks between the cryosphere and climate are not yet well quantified, either within the Arctic or globally. This lends considerable uncertainty to predictions of how much and how fast the cryosphere and the Arctic environment will change.

How the Arctic cryosphere is changing

The extent and duration of snow cover and sea ice have decreased across the Arctic. Temperatures in the permafrost have risen by up to 2 °C. The southern limit of permafrost has moved northward in Russia and Canada.

Key finding 3

The largest and most permanent bodies of ice in the Arctic – multi-year sea ice, mountain glaciers, ice caps and the Greenland Ice Sheet – have all been declining faster since 2000 than they did in the previous decade.

Key finding 4

Model projections reported by the Intergovernmental Panel on Climate Change (IPCC) in 2007 underestimated the rates of change now observed in sea ice.

Key finding 5

The extent and duration of snow cover have decreased throughout the Arctic. The Arctic land area covered by snow in early summer has reduced by 18% since 1966. Coastal areas of Alaska and northern Fennoscandia have seen the strongest decreases in the number of days with snow cover. The change is largely caused by snow melting earlier in the spring. Snow depth has decreased in some areas such as the North American Arctic, but has increased in others such as northern Russia.

Permafrost – permanently frozen ground – underlies most of the Arctic land area and extends under parts of the Arctic Ocean. Temperatures in the permafrost have risen by up to 2 °C over the past two to three decades, particularly in colder sites (typical permafrost temperatures range from -16 °C to just below 0 °C, depending on the location). The depth of soil above the permafrost that seasonally thaws each year has increased in Scandinavia, Arctic Russia west of the Urals, and inland Alaska. The southern limit of the permafrost retreated northward by 30 to 80 km in Russia between 1970 and 2005, and by 130 km during the past 50 years in Quebec.

Ice cover on lakes and rivers in the Northern Hemisphere is breaking-up earlier than previously observed. Studies of sediments in High Arctic lakes indicate that the duration of ice cover on some lakes has declined significantly over the past 100 years. The rates of change in lake and river ice conditions vary across the Arctic, although there are few long-term systematic observations.

Large bodies of ice are melting faster

Net loss of mass from the Greenland Ice Sheet has increased from an estimated 50 Gt per year (50 000 000 000 metric tonnes per year) in the period 1995–2000 to ~200 Gt per year in the period 2004–2008. The current loss (~200 Gt per year) represents enough water to supply more than one billion city-dwellers.

Nearly all glaciers and ice caps in the Arctic have shrunk over the past 100 years. The rate of ice loss increased over the past decade in most regions, but especially in Arctic Canada and southern Alaska. Total loss of ice from glaciers and smaller ice caps in the Arctic probably exceeded 150 Gt per year in the past decade, similar to the estimated amount being lost from the Greenland Ice Sheet.

Arctic sea-ice decline has been faster during the past ten years than in the previous 20 years. This decline in sea-ice extent is faster than projected by the models used in the IPCC's Fourth Assessment Report in 2007. The area of sea ice persisting in summer (polar pack ice) has been at or near record low levels every year since 2001. It is now about one third smaller than the average summer sea-ice cover from 1979 to 2000. New observations reveal that average sea-ice thickness is decreasing and the sea-ice cover is now dominated by younger, thinner ice.

More change is expected

Maximum snow depth is expected to increase over many areas by 2050, with greatest increases over Siberia. Despite this, average snow cover duration is projected to decline by up to 20% by 2050.

Key finding 6

The Arctic Ocean is projected to become nearly ice-free in summer within this century, likely within the next thirty to forty years.

Key finding 7

Average Arctic autumn-winter temperatures are projected to increase by between 3 and 6 °C by 2080, even using scenarios in which greenhouse gas emissions are projected to be lower than they have been for the past ten years.

The climate models used for SWIPA do not include possible feedback effects within the cryosphere system that may release additional stores of greenhouse gases from Arctic environments.

Arctic snowfall and rain are projected to increase in all seasons, but mostly in winter. Despite this, Arctic landscapes are generally expected to dry out more during summer. This is because higher air temperatures mean that more water evaporates, snow melt finishes earlier, and water flow regimes are altered.

With increasing snowfall, all projections show maximum snow depth during winter increasing over many areas. The greatest increases (15–30% by 2050) are expected in Siberia. Even so, snow will tend to lie on the ground for

10–20% less time each year over most of the Arctic, due to earlier melting in spring.

Models project continued thawing of permafrost.

Projections show that sea-ice thickness and summer sea-ice extent will continue to decline in the coming decades, although considerable variation from year to year will remain. A nearly ice-free summer is now considered likely for the Arctic Ocean by mid-century. This means there will no longer be any thick multi-year ice consistently present.

Climate model projections show a 10–30% reduction in the mass of mountain glaciers and ice caps by the end of the century.

The Greenland Ice Sheet is expected to melt faster than it is melting now, but no current models can predict exactly how this and other land-based ice masses in the Arctic will respond to projected changes in the climate. This is because ice dynamics and complex interactions between ocean, snow, ice and the atmosphere are not fully understood.

How these changes affect Arctic ecosystems and people

Changes in the cryosphere cause fundamental changes to the characteristics of Arctic ecosystems and in some cases loss of entire habitats. This has consequences for people who receive benefits from Arctic ecosystems.

Key finding 8

The observed and expected future changes to the Arctic cryosphere impact Arctic society on many levels. There are challenges, particularly for local communities and traditional ways of life. There are also new opportunities.

Key finding 9

Changes in the cryosphere cause fundamental changes in Arctic ecosystems

Changes in the amount of snow and the structure of the snowpack affect soils, plants and animals. Some species, such as pink-footed goose, benefit from less snow cover in spring. But animals grazing through snow, such as reindeer/caribou, suffer if winter rainfall creates an ice-crust over the snow. This is already happening more often in northern Canada and Scandinavia.

Less snow and faster melting are causing summer drought in forests, wetlands, and lakes supplied by snow melt. Thawing permafrost is also causing wetlands in some areas to drain and dry out, while creating new wetlands elsewhere.

The loss of ice cover over rivers, lakes and seas is changing animal and plant communities in the water.

The loss of large areas of sea ice represents devastating habitat loss for some ice-adapted species, including polar bear, seals, walrus, narwhal and

some microbial communities. Many animals, including bowhead whales, depend on tiny crustaceans that thrive near the sea ice. This food source is changing as the ice edge recedes.

These changes to ecosystems directly affect supplies of water, fish, timber, traditional/local foods and grazing land used by Arctic people. For example, it has been suggested that stocks of some sub-Arctic and Arctic-adapted fish species, including commercially important species, could change as sea ice recedes. Uncertainty about changing supplies of living natural resources makes it difficult to plan for the future.

Forestry may benefit from thawing permafrost in areas where there is enough water for trees to grow, but insect pests are increasingly causing problems. Some hunted animals, such as seals and walruses, are declining in numbers as ice conditions change. Others are moving to new locations, so hunters have to travel further to reach them.

Transport options and access to resources are radically changed by differences in the distribution and seasonal occurrence of snow, water, ice and permafrost in the Arctic. This affects both daily living and commercial activities.

Key finding 10

Arctic infrastructure faces increased risks of damage due to changes in the cryosphere, particularly the loss of permafrost and land-fast sea ice.

Key finding 11

Cryospheric change affects Arctic livelihoods and living conditions

Access to northern areas via the sea is increasing during the summer as sea ice disappears; allowing increased shipping and industrial activity. Offshore oil and gas activities will benefit from a longer open water season, although threats from icebergs may increase due to increased iceberg production. The International Maritime Organization is devising new mandatory guidelines for ships operating in ice-covered waters. Sea-ice decline creates challenges for local residents who use the ice as a platform for travel and hunting; these challenges may include travelling farther over uncertain ice conditions and increased hazards.

On land, access to many areas is becoming more difficult as ice roads melt earlier and freeze later and as permafrost degrades. Industrial operations reliant on ice roads will need to concentrate heavy load transport into the coldest part of the year. Shorter seasons where ice and snow roads can be used severely impact communities that rely on land transport of goods to maintain reasonable retail costs and ensure economic viability, particularly in northern Canada and Russia. Some land areas become more accessible for mining as glaciers and ice caps recede.

Thawing permafrost is causing increased deformation of buildings, roads, runways and other man-made structures in some areas, although poor design in the past is a contributing factor. New design methods are being developed that consider the likelihood of environmental change. Buildings and other infrastructure are at risk from heavier snow loads and floods caused by the release of ice jams in rivers or sudden emptying of glacial lakes.

Two-thirds of the Arctic coastline is held together and protected by ice. When land-fast sea ice melts earlier and permafrost thaws, rapid erosion can occur. Along the coasts bordering the Laptev and Beaufort seas, coastal retreat rates of more than two metres per year have been recorded. A number of Inuit villages in Alaska are preparing to relocate in response to the encroaching sea.

In the short term, increased glacier melt creates new opportunities for hydroelectricity generation. This has potential benefits for industry. In the longer term, the volume of meltwater will decrease as glaciers shrink, potentially affecting electricity production.

Melting ice and snow release contaminants that have been stored for many years, allowing the contaminants to re-enter the environment. Exposure of people and top predators to contaminants that accumulate in food chains could further increase.

Increased access to the Arctic creates new economic opportunities. Cruise ship tourism is increasing. More people are coming to witness the effects of climate change on Arctic glaciers, for example at the Ilulissat Icefjord in Greenland. Increased tourism may challenge lifestyles and services in local communities as well as increase the demand for effective infrastructure (e.g., air services, marine navigation aids, and other safety measures). Loss of Arctic wildlife and change of scenery may adversely affect the tourist industry in the long term.

Cryospheric change combined with rapid development creates opportunities and challenges for Arctic residents. Traditional knowledge can help to detect change and adapt to it. While traditional knowledge continues to evolve, it is a challenge to ensure that this knowledge is being passed on to younger generations as lifestyles change. Some aspects of traditional knowledge become less applicable as the cryosphere and other components of the Arctic system change even more rapidly and become less predictable.

Why changes in the Arctic matter globally

Changes in the Arctic cryosphere have impacts on global climate and sea level

When highly reflective snow and ice surfaces melt away, they reveal darker land or ocean surfaces that absorb more of the sun's energy. The result is enhanced warming of the Earth's surface and the air above it. There is evidence that this is happening over the Arctic Ocean as the sea ice retreats, as well as on land as snow melts earlier.

Overall emissions of methane and carbon dioxide from the Arctic could

Loss of ice and snow in the Arctic enhances climate warming by increasing absorption of the sun's energy at the surface of the planet. It could also dramatically increase emissions of carbon dioxide and methane and change large-scale ocean currents. The combined outcome of these effects is not yet known.

Key finding 12

Arctic glaciers, ice caps and the Greenland Ice Sheet contributed over 40% of the global sea level rise of around 3 mm per year observed between 2003 and 2008. In the future, global sea level is projected to rise by 0.9–1.6 m by 2100 and Arctic ice loss will make a substantial contribution to this.

Key finding 13

increase due to warming of soils and freshwater systems, and thawing of ancient frozen soil beneath the seabed.

The combined outcome of these effects on global climate is not yet known.

All the main sources of freshwater entering the Arctic Ocean are increasing – river discharge, rain/snow, and melting glaciers, ice caps, and the Greenland Ice Sheet. Recent calculations estimate that an extra 7700 km^3 of freshwater – equivalent to one metre of water over the entire land surface of Australia – has been added to the Arctic Ocean in recent years. There is a risk that this could alter large-scale ocean currents that affect climate on a continental scale.

Melting glaciers and ice sheets worldwide have become the biggest contributor to global sea level rise. Arctic glaciers, ice caps, and the Greenland Ice Sheet contributed 1.3 mm – over 40% – of the total 3.1 mm global sea level rise observed every year between 2003 and 2008. These contributions from the Arctic to global sea level rise are much greater than previously measured.

High uncertainty surrounds estimates of future global sea level. Latest models predict a rise of 0.9 to 1.6 m above

the 1990 level by 2100, with Arctic ice making a significant contribution.

Changes in the Arctic cryosphere affect global society

Sea level rise is one of the most serious societal impacts of cryospheric change. Higher average sea level and more damaging storm surges will directly affect millions of people in low-lying coastal flood plains. Sea level rise increases the risk of inundation in coastal cities such as Shanghai and New York.

On the other hand, global economic activity may benefit from cryospheric changes in the Arctic. For example, opening transpolar sea routes across the Arctic Ocean will reduce the distance for ships travelling between Europe and the Pacific by 40% compared to current routes. This could reduce emissions and energy use.

Some unique Arctic species, such as the narwhal, face particular threats as the cryosphere changes. The decline of cryospheric habitats such as sea ice and wetlands over permafrost will impact on migratory species of mammals and birds from elsewhere in the world. These adverse effects on biodiversity are of global concern.

What should be done

Everyone who lives, works or does business in the Arctic will need to adapt to changes in the cryosphere. Adaptation also requires leadership from governments and international bodies, and increased investment in infrastructure.

Key finding 14

Adaptation is urgent and needed at all levels

Cryospheric change affects people at the local level first, and local communities will need to devise strategies to cope with emerging risks.

At national and regional levels, adaptation requires leadership from governments and international bodies to establish new laws and regulations. For example, new fishing regulations will be required as fish stocks change. New standards will need to be developed for construction, particularly in areas affected by thawing permafrost.

Governments will need to invest in transport networks to cope with the shorter ice road season. Search and rescue operations will need to be enhanced to respond to increasing traffic and risks at sea, and accurate forecasts of weather and sea conditions are required to ensure travel safety.

Arctic communities are resilient and will actively respond to cryospheric change. However, rapid rates of change may outpace adaptation capacity. Knowledge and research are needed to foresee how living conditions are likely to change, and to evaluate possible adaptation options. Concerns of indigenous peoples need particular attention in this regard.

Changes in the cryosphere are not the only driver of change in the Arctic. Cryospheric change and climate change occur in the context of societal change, which may be even more challenging. The combined effects of societal, climatic and cryospheric change must be taken into account in adaptation strategies.

Cutting greenhouse gas emissions globally is urgent

Climate change represents an urgent and potentially irreversible threat to human societies. Global climate modeling studies show that deep and immediate

There remains a great deal of uncertainty about how fast the Arctic cryosphere will change in the future and what the ultimate impacts of the changes will be. Interactions ('feedbacks') between elements of the cryosphere and climate system are particularly uncertain. Concerted monitoring and research is needed to reduce this uncertainty.

Key finding 15

cuts in global greenhouse gas emissions are required to hold the increase in global average temperatures below 2 °C above pre-industrial levels. Combating human-induced climate change is an urgent common challenge for the international community, requiring immediate global action and international commitment.

Following the ACIA report, published in 2005, Ministers of the Arctic Council acknowledged that "*timely, measured and concerted action is needed to address global emissions.*" They endorsed a number of policy recommendations for reducing greenhouse gas emissions, including to "*Adopt ... strategies ... [to] address net greenhouse gas emissions and limit them in the long term to levels consistent with the ultimate objective of the UNFCCC* [United Nations Framework Convention on Climate Change]."

The key findings of the SWIPA assessment, especially the rapid and accelerated rates of change in Arctic cryosphere conditions, emphasize the need for greater urgency in taking these actions.

Uncertainty can be reduced by further research

Current monitoring, research and model results provide high confidence that significant changes are occurring in the Arctic cryosphere and that these changes will continue in the future. Some of the observed changes align with expectations but one major component of the cryosphere (sea ice) has reacted faster than anticipated just five years ago. Even so, substantial uncertainty remains, particularly concerning the future timing of changes, and the effects of interactions (feedbacks) between components of the cryosphere and climate system.

To reduce the uncertainty in future assessments, more robust observational networks are needed. Satellites and airborne measurements have improved the ability to observe some elements of the Arctic cryosphere such as sea-ice extent and snow cover. Monitoring of other key elements of the cryosphere, notably sea-ice thickness, snow depth, permafrost and glaciers, requires surface-based observations.

Many surface-based snow, freshwater ice, and precipitation gauge networks have diminished or have been completely lost, and sites for measuring sea ice, land ice, and physical properties of snow are sparse. Observational networks need to be expanded to provide a robust set of cryospheric data for monitoring, model improvement and satellite product validation.

The biggest unanswered questions identified by this report are:

- What will happen to the Arctic Ocean and its ecosystems as freshwater is added by melting ice and increased river flow?
- How quickly could the Greenland Ice Sheet melt?
- How will changes in the Arctic cryosphere affect the global climate?
- How will the changes affect Arctic societies and economies?

Answering some of these questions requires improved monitoring networks. A better understanding of the complex interactions between the physical, chemical and biological environments in the Arctic is needed. There is a lack of systematically collected information on the effects of cryospheric change on human society.

Communicating about cryospheric change and its effects

The SWIPA assessment documents the importance of climate-induced changes in Arctic snow, water and ice conditions and the profound implications for the local, regional, and global society.

Active communication of this new knowledge, to enhance global, national, and local awareness, will help to ensure that the SWIPA assessment generates benefits for people in the Arctic.

A co-ordinated response to cryospheric change

The combined effects of the changing cryosphere, climate change, and rapid development in the Arctic will create political challenges for Arctic societies, as well as the global community. Traditional livelihoods are most vulnerable to changes in the cryosphere. There is a need for co-operation and co-ordinated effort at all levels, to respond to change and increase the resilience of Arctic ecosystems and societies.

Recommendations

Based on the SWIPA key findings, the AMAP Working Group have agreed to the following recommendations:

Adaptation

Members of the Arctic Council and governments at all levels in the Arctic should work to:

- Develop regional-scale assessments of cryospheric change and the associated risks.

- Develop and implement Arctic adaptation strategies appropriate to the scale and character of anticipated changes. Such strategies must take account of other relevant drivers of change.

- Ensure that standards for environmental management are in place, or can be adapted, to take account of cryospheric change. Develop regulations where necessary.

- Upgrade the capacity for search and rescue operations and environmental hazard responses.

- Facilitate measures to increase the accuracy of forecasting for ice, weather, and sea conditions, and make forecasts accessible to all Arctic residents and organizations.

Mitigation

International negotiations to reduce global greenhouse gas emissions should be pursued as a matter of urgency.

Member States of the Arctic Council should increase their leadership role in this process.

Observation

Arctic countries and international organizations should:

- Improve and expand systematic, comprehensive surface-based monitoring of the cryosphere.

- Maintain and support development of remote sensing methods for observing the cryosphere.

- Develop and enhance systems to observe the cascading effects of cryospheric change on ecosystems and human society.

- Expand research into processes that are important for modeling the cryosphere, to reduce uncertainty in predicting cryospheric change. In particular, improvements are needed in modeling permafrost dynamics, snow-vegetation interactions, and mass loss from glaciers, ice caps, and the Greenland Ice Sheet.

Outreach

The Members and Observers of the Arctic Council should individually and collectively inform and educate Arctic societies and the global society about the changes in the Arctic linked to climate change, and how they affect people locally, regionally and globally.

Policy Needs

Governments and institutions at all levels should increase co-operation and co-ordinate efforts to respond to the challenges and opportunities associated with cryospheric change.

The Arctic Council should conduct an integrated assessment of the combined impacts of change in the Arctic, focused on how to minimize environmental damage and enhance human well-being.

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 synthesize and evaluate 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 1.1). 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 1.2) 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.

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 snow cover, permafrost, lake and river ice, mountain glaciers and ice caps, the Greenland Ice Sheet, and sea ice conditions, 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 the benchmark for the SWIPA assessment, which focuses on recent change in the Arctic cryosphere and the effects of observed and projected

Figure 1.1. The Arctic, as defined by AMAP and as used in the present assessment.
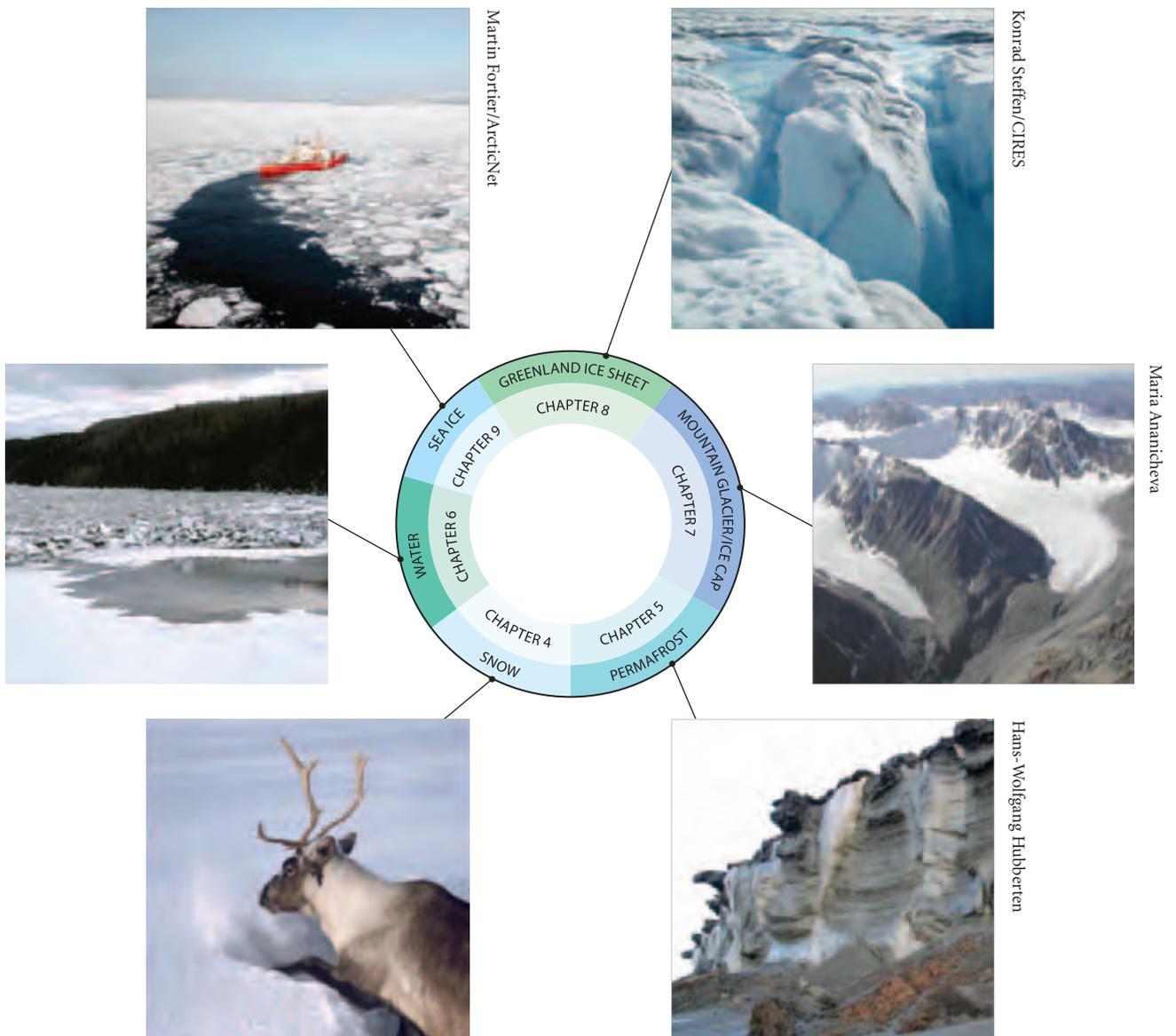
Figure 1.2. The various components of the Arctic cryosphere.

change. The SWIPA assessment is thus an update and extension of the ACIA findings on the consequences of change in the Arctic cryosphere component of the global climate system. Figure 1.3 presents a conceptual illustration of the various components of the SWIPA assessment and the links between them.

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
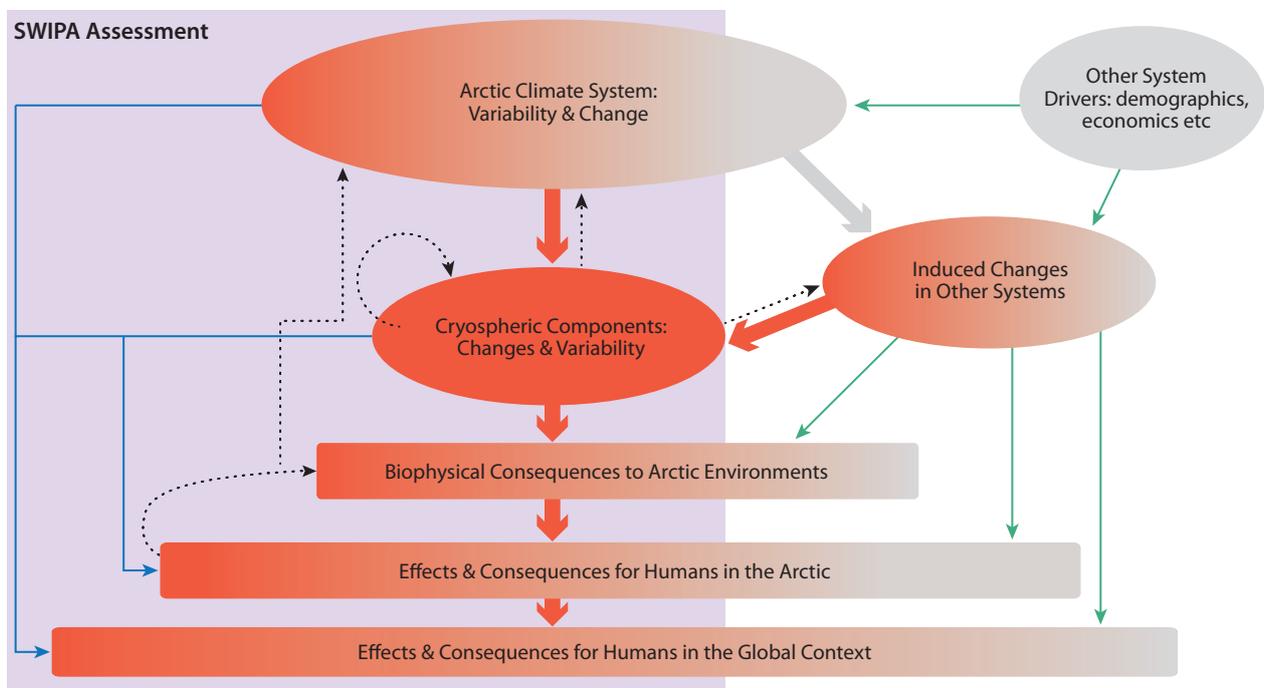
Figure 1.3. The Arctic cryosphere and linkages to various levels of effects or consequences of changes. The SWIPA assessment focuses on the red components highlighted in the various elements of the overall system and the cascade of effects shown as thick arrows. Several points emerge from the diagram: (i) Arctic climate change induces change in systems other than just the cryosphere itself (blue arrows); (ii) other drivers of the overall system induce changes in various elements (green 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 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.

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.

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 thawing 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.

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.

1.5. What SWIPA does and does not cover

Cryospheric change and variability is fundamentally linked to climate change (see Boxes 1.1 to 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 the benchmark for the SWIPA assessment, and an assessment against which the new information presented in the SWIPA assessment can be compared.

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.

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 either 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 on Arctic cryospheric changes driven by underlying climate changes rather than on 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 1.4). 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 to 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 1.5 showing historical forcing over time, where the forcing influence is normalized in terms of W/m². Volcanoes have a cooling influence of up to -5 W/m² 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/m² by 2000. Sulfate (SO_4) is an aerosol with a cooling influence of 0.6 W/m² by 2000. Solar forcing has decadal and centennial variability, but its influence over the previous 200 years is below 0.4 W/m². 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 1.4. Near-surface air temperature anomaly multiyear composites for 2002–2009. Anomalies are relative to the 1951–1980 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.

Surface temperature anomalies 2002–2009 vs 1951–1980

-1.5
-0.5
0.5
1.0
1.5
2.0
3.5
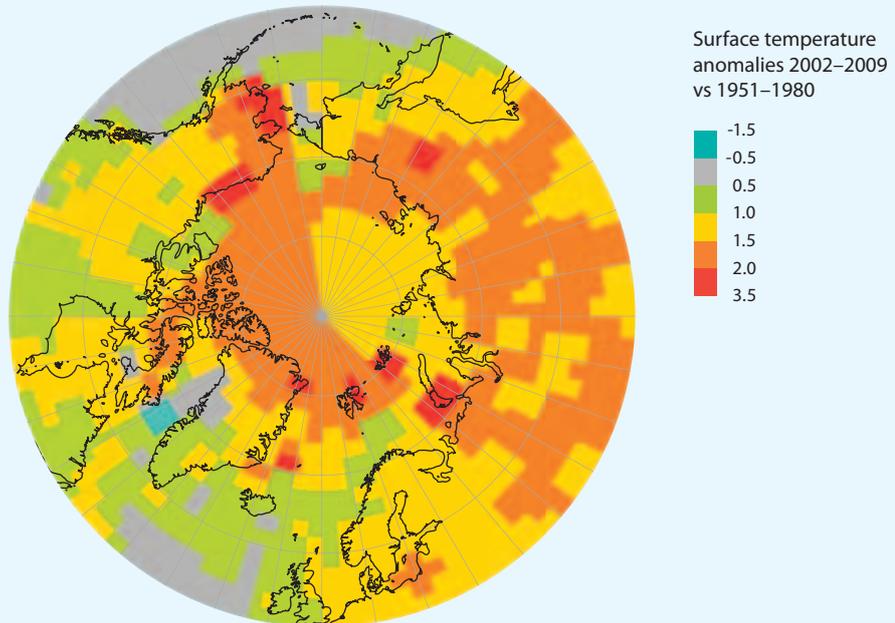
Figure 1.5. Relative forcing of 20th century northern hemispheric temperature increases. Redrawn after Crowley (2000). Note scale differences in y-axes.
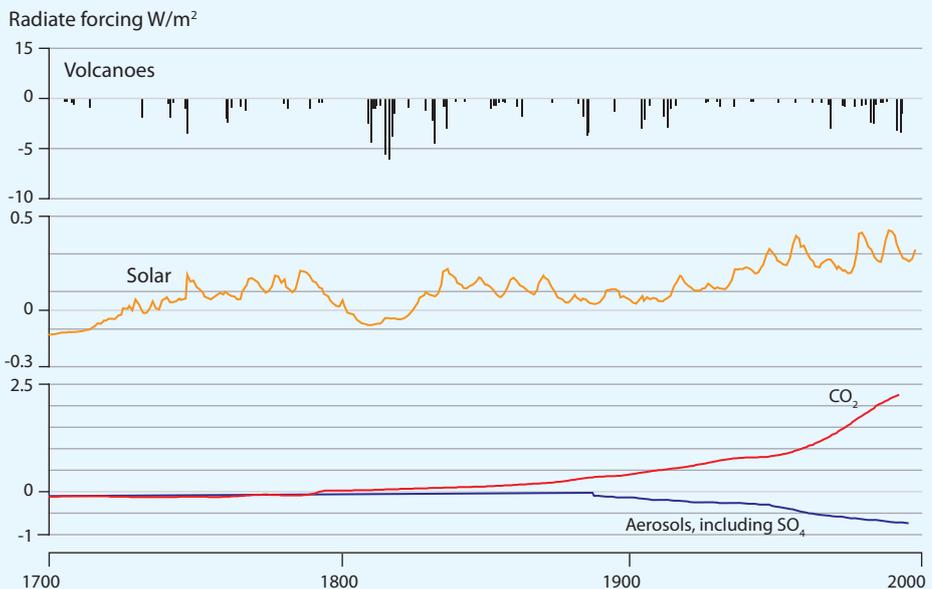
All parts of the Arctic cryosphere are a product of and subject to climatic conditions, but whereas seasonal phenomena (such as seasonal snow cover and seasonal ice on rivers and lakes) respond to specific weather conditions and rapidly changing climatic conditions, other parts of the cryosphere (such 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 was presented to the Arctic Council at its meeting in May 2011 (AMAP, 2011a).

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.

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 those of the IPCC) were used throughout this report (see Figure 1.6).

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 strict and independent peer review process was established by the AMAP Working Group to secure and document the integrity of the process (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 (AMAP, 2012). A SWIPA Summary for Policymakers produced under the AMAP Working Group and presented to the Arctic Council Ministers at their meeting in
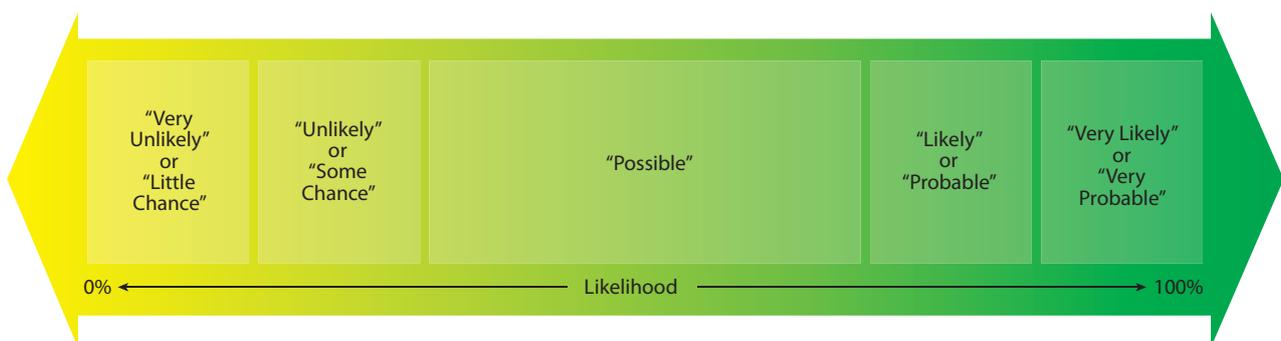
Figure 1.6. Five-tier lexicon describing the likelihood of expected change.

Nuuk, Greenland, 2011 includes policy-relevant scientific recommendations (AMAP, 2011b). The scientific SWIPA assessment report (the present report) provides the validated scientific basis for all statements made in the overview 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 describes 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

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

Contents

Key Findings

- 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 an influence of sea-ice reduction, 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, where wind-driven export of older sea ice appears to have preconditioned the Arctic Ocean for its recent rapid ice loss.

- 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.

- 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.

- 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.

- 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.

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 was 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.

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- 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 (Kaufman 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.
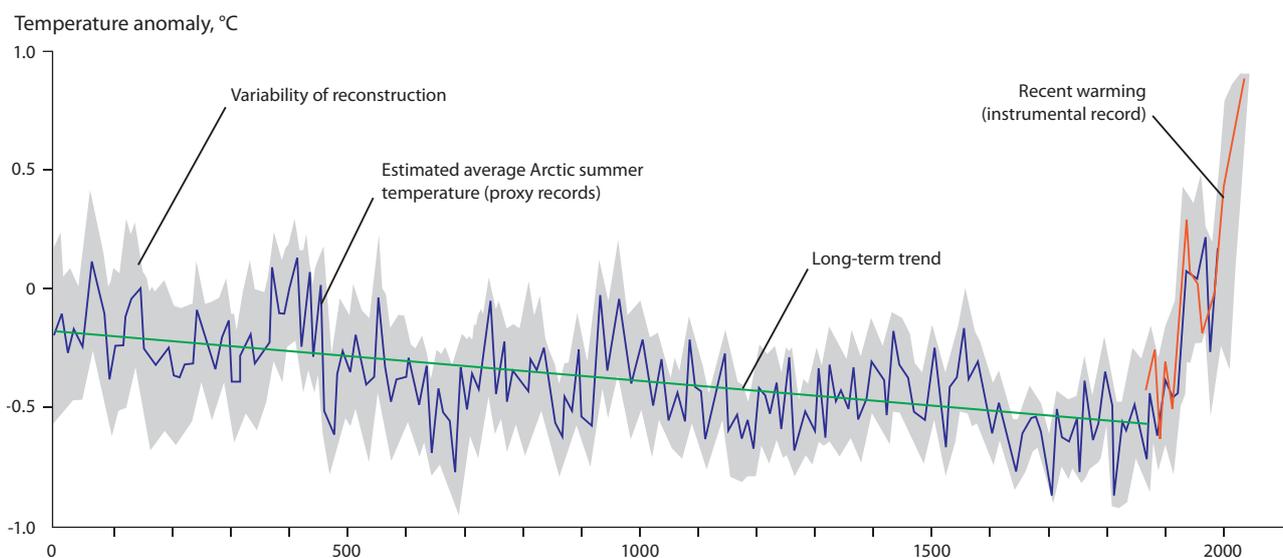
Figure 2.1. Estimated Arctic average summer air temperature anomalies for the past 2000 years (relative to the 1961 to 1990 mean), based on proxy records from lake sediments, ice cores, and tree rings. The shaded area represents variability among the 23 sites used for the reconstruction. Source: Kaufman 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) through 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 2004, 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 snow cover, glacier mass and sea ice discussed in Chapters 4, 7, 8 and 9. 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.

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. The seasonal variation of the warming is consistent with greenhouse-driven changes projected by global climate models (see Chapter 3).
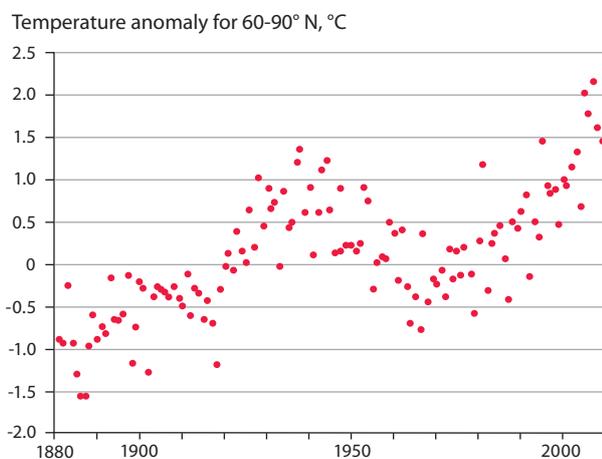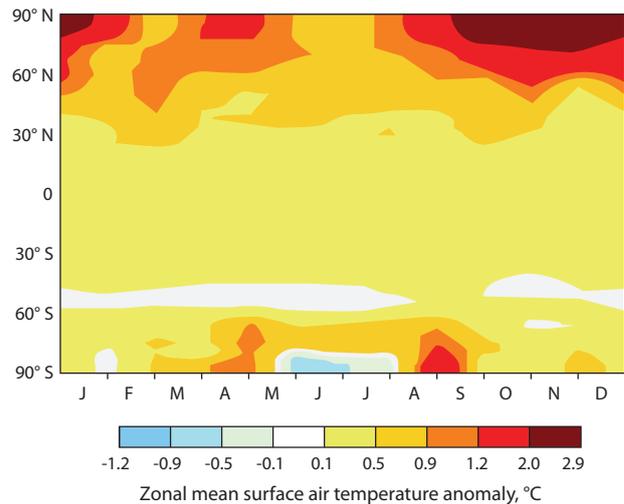
Figure 2.3. Monthly surface air temperature anomalies averaged over the period 2001 to 2009 (relative to the mean for 1951 to 2000) shown as a function of latitude. Source: NASA Goddard Institute for Space Studies (http://data.giss.nasa.gov/gistemp).

The 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 and seasonal (Figure 2.4) 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 2.4) 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 2.4 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 2.4) 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 Arctic Oscillation (AO), which drives temperature anomalies from eastern Canada across the North Atlantic to northern Eurasia (Thompson and Wallace, 2000), and the Pacific Decadal Oscillation (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

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 Climate Data Center.

attributed by various researchers (e.g., Comiso, 2003; Overland et al., 2008) to a predominantly positive phase of the AO (Figure 2.5). 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 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
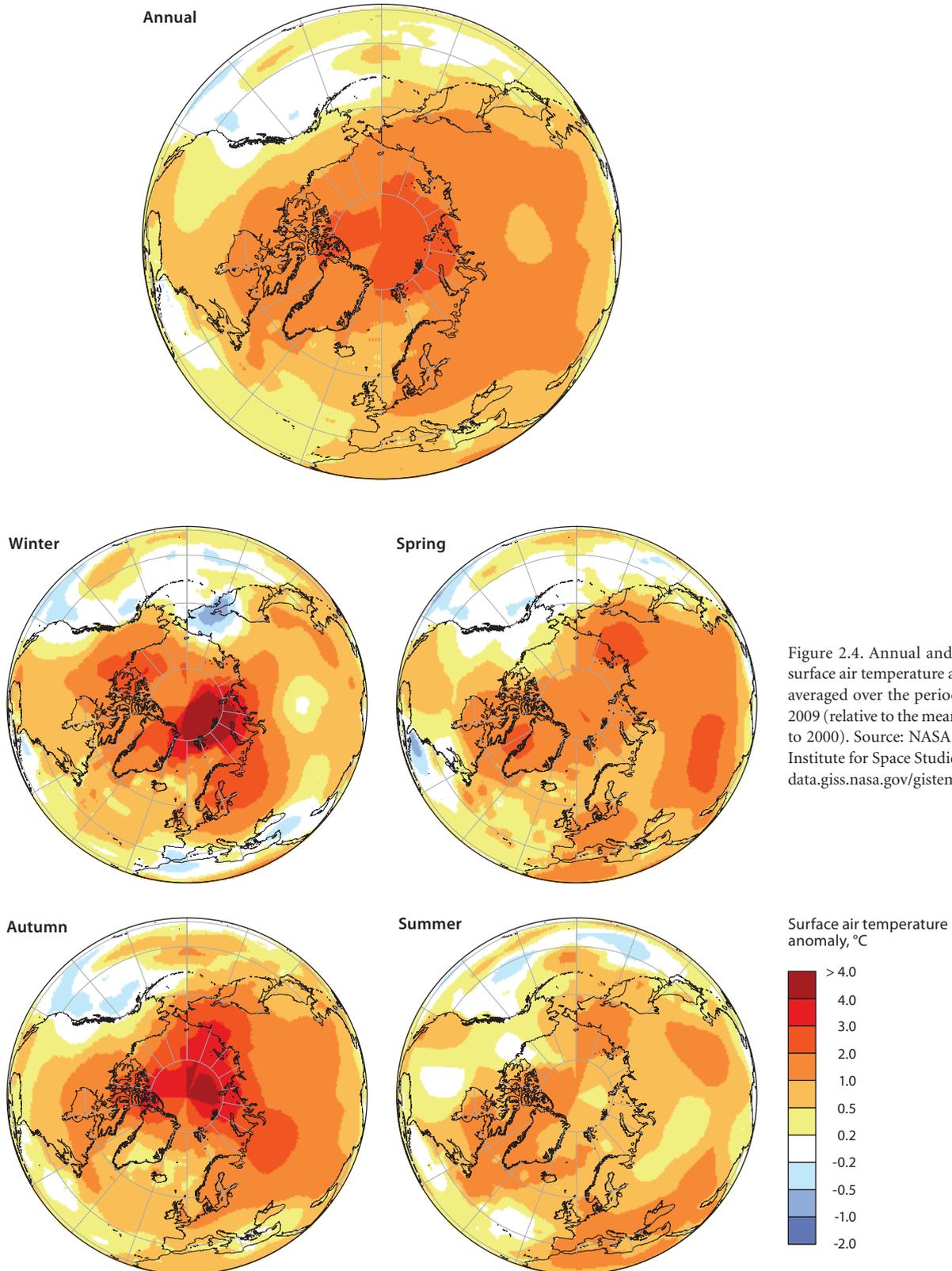
Figure 2.4. Annual and seasonal surface air temperature anomalies averaged over the period 2005 to 2009 (relative to the mean for 1951 to 2000). Source: NASA Goddard Institute for Space Studies (http://data.giss.nasa.gov/gistemp).

AO index (standardized 3-month running mean)
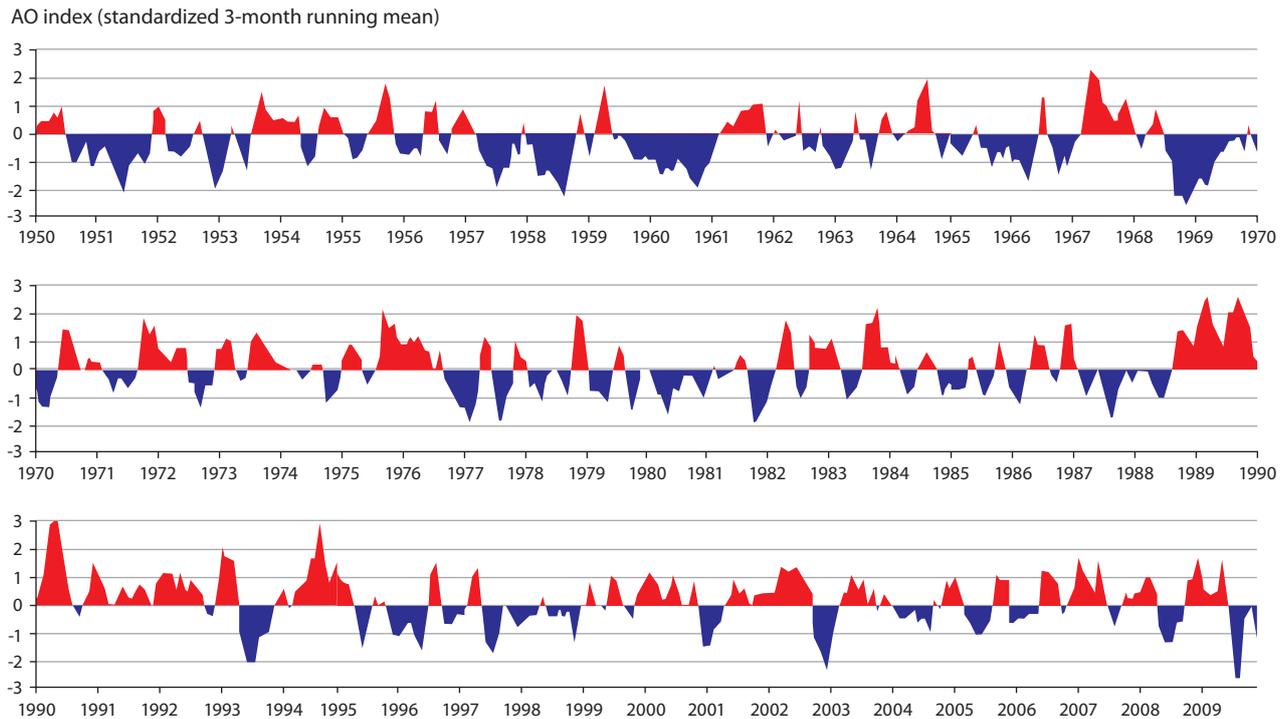
Figure 2.5. AO index, plotted as a standardized 30-month running mean for the period 1950 to 2009. Source: NOAA/Climate Prediction Center.

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 2.6 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 couplet of temperature anomalies of opposite sign in the winter pattern (see Figure 2.4) 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 2.7), 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 2.7 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).

Monthly PDO index
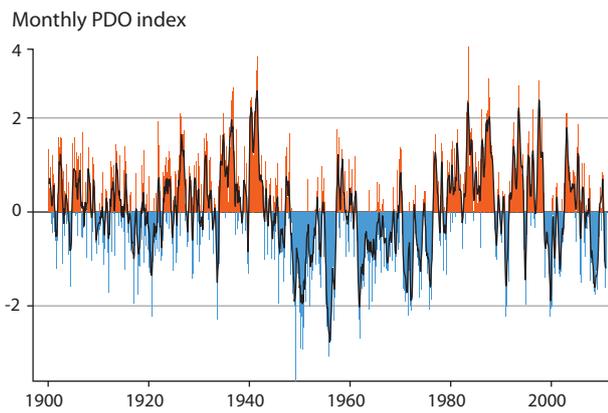
Figure 2.6. Pacific Decadal Oscillation index, expressed in terms of monthly values and a five-month running mean (black line). Source: Joint Institute for the Study of the Atmosphere and Ocean, NOAA/University of Washington (http://jisao.washington.edu/pdo).

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.
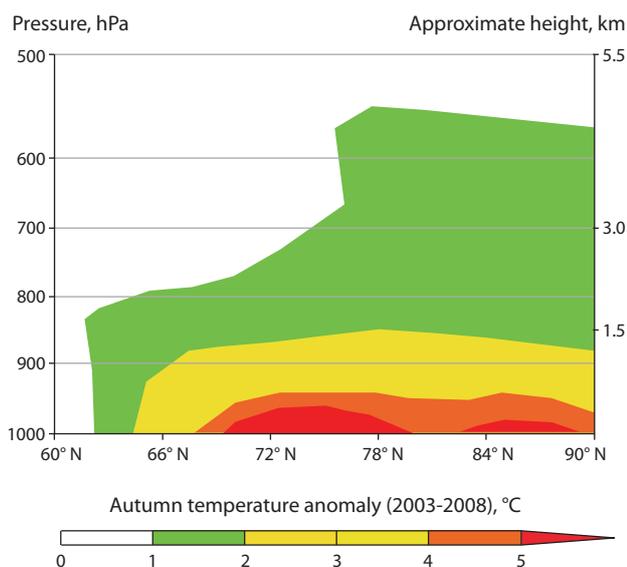
Figure 2.7. Latitude-height cross-section of pan-Arctic autumn (October / November / December) temperature anomalies for 2003-2008 (relative to the 1858-96 base period). Source: J. Overland, NOAA; based on the National Centers for Environmental Prediction / National Center for Atmospheric Research (NCAR/NCEP) re-analysis.

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 11.5, 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 2.8 for monthly precipitation averaged over the land areas north of 55° N. The data show a strong 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 2.9). 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 2.10). 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, 2003).
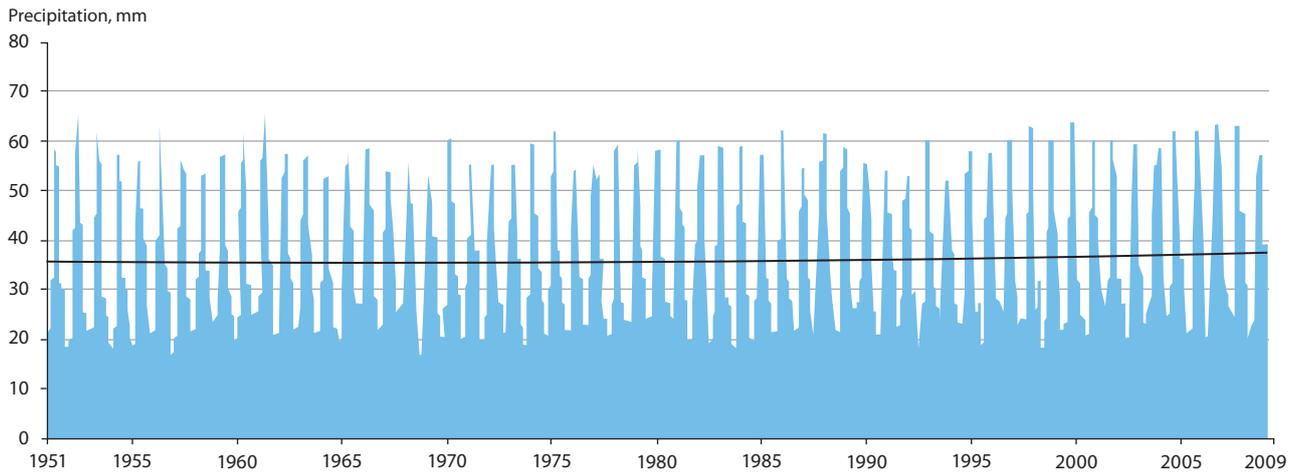
Precipitation, mm

Figure 2.8. Monthly precipitation averaged over land areas north of 55° N for the period 1951 to 2009. The solid black line shows a least-squares polynomial fit. Source: Global Precipitation Climatology Center / World Meteorological Organization / Deutscher Wetterdienst (Courtesy of B. Rudolf).

Annual precipitation, mm
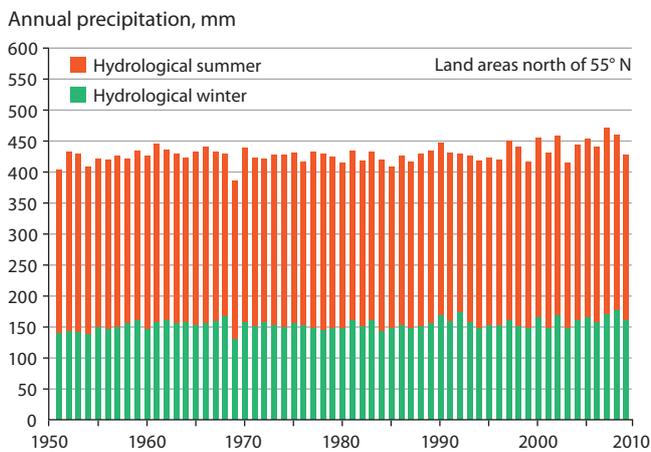
Figure 2.9. Annual precipitation for the period 1951 to 2009 averaged over land areas north of 55° N for hydrological winter (October to March) and hydrological summer (April to September). Source: Global Precipitation Climatology Center / World Meteorological Organization / Deutscher Wetterdienst (Courtesy B. Rudolf).

Precipitation anomaly, mm
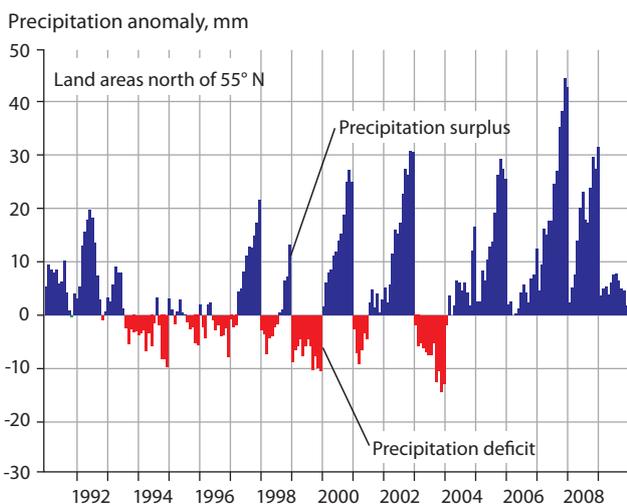
Figure 2.10. Monthly-accumulated anomalies of precipitation for the period 1991 to 2009 (relative to the corresponding monthly means for 1951 to 2000) averaged over land areas north of 55° N. Source: Global Precipitation Climatology Center / World Meteorological Organization / Deutscher Wetterdienst (Courtesy of B. Rudolf).

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 2.11). For Eurasia, the discharge of the largest rivers has increased by about 10% since 1935, despite the large 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 2.8 and 2.9). 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

Figure 2.11. 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 red dots. Source: Richter-Menge and Overland (2009). Courtesy of A. Shiklomanov.
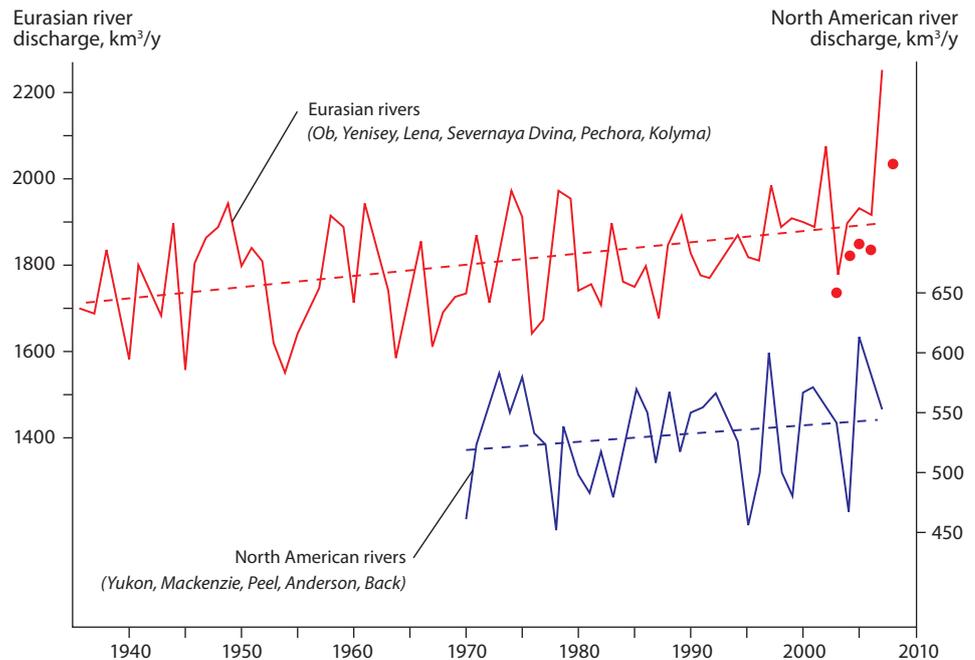
related to retreating sea ice, the role of storminess in itself can be difficult to unravel. Nevertheless, it is apparent from the absence of a comprehensive (pan-Arctic) evaluation of recent variations in storminess that there is a need for systematic assessments of storminess in the Arctic. Such an 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),
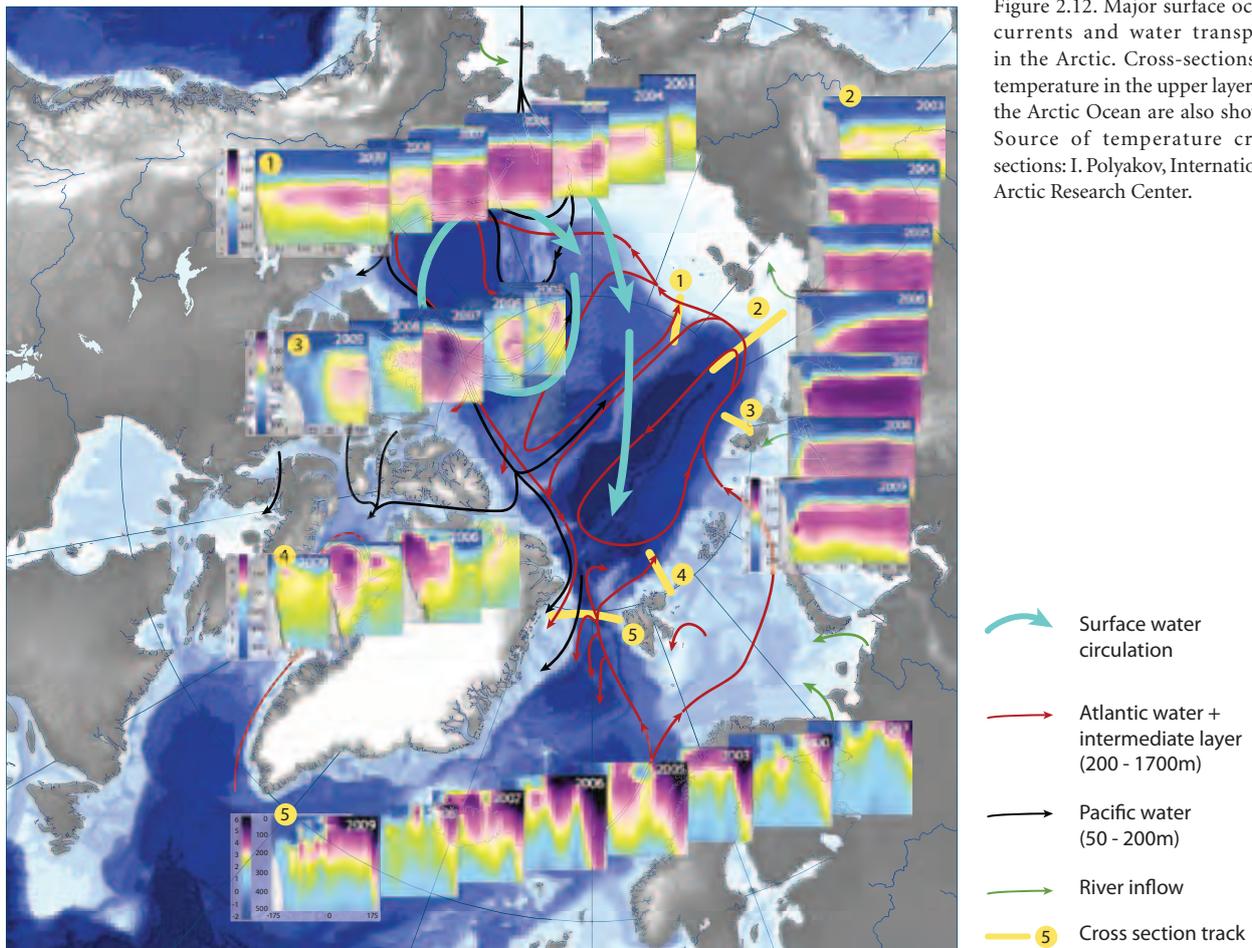
Figure 2.12. Major surface ocean currents and water transport in the Arctic. Cross-sections of temperature in the upper layers of the Arctic Ocean are also shown. Source of temperature cross sections: I. Polyakov, International Arctic Research Center.

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 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 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 2.12), consistent with the mean pattern of Arctic Ocean currents. 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 2.12, 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

Box 2.1. Are changes in the Arctic affecting mid-latitude weather?

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 2.13a). This pattern broke down in December 2009 (Figure 2.13b); 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 2.14 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.
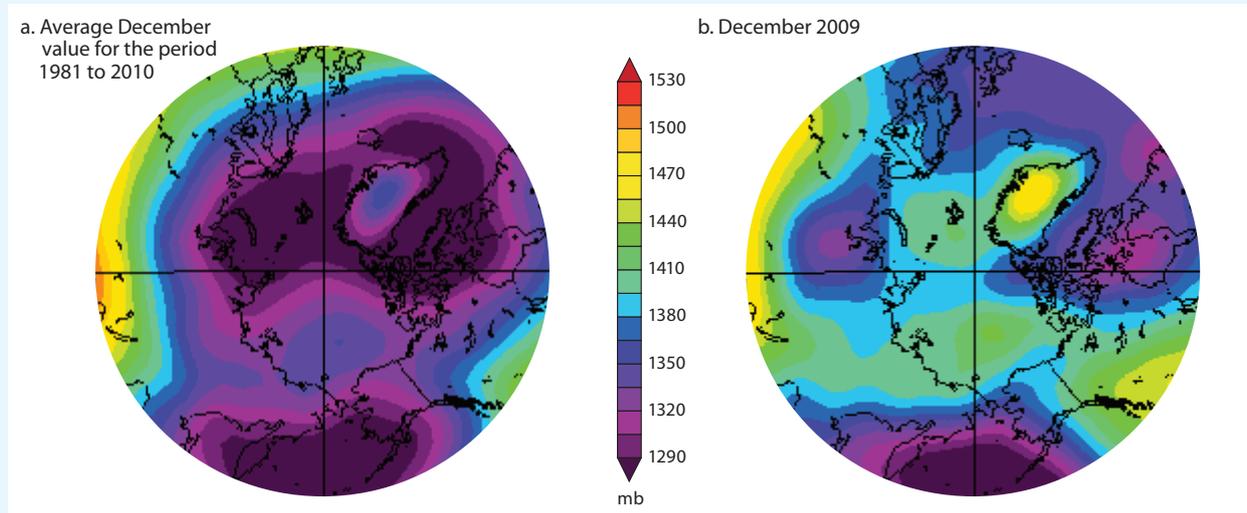
Figure 2.13. Arctic atmospheric pressure fields: (a) average 850 mb geopotential height values for December from 1981 to 2010, (b) with unusual 850 mb geopotential height maximums observed over the Arctic in December 2009. Figures from NOAA/ESRL Physical Sciences Division.
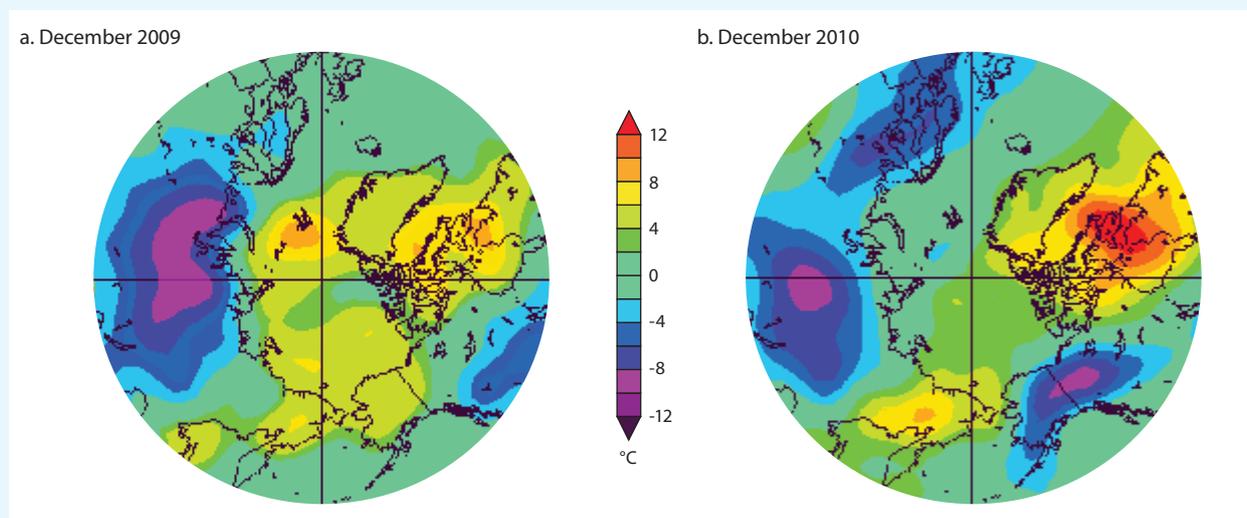
Figure 2.14. The Warm Arctic-Cold Continents Climate Pattern for (a) December 2009 and (b) December 2010. Plots show anomalies or deviations from the normal 1000 mb air temperature values which were observed from 1981 to 2010. Data are from the NCEP-NCAR Reanalysis through the NOAA/ESRL Physical Sciences Division.

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 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 recent occurrence of an Arctic influence on mid-latitude weather is reported in Box 2.1.

The results presented in this chapter also point to various observational needs. These are discussed in more detail in Chapter 11, Section 11.5. 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 11.5, 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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3. Climate Model Projections for the Arctic

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Contents

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?

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)

a. Observed

b. IPCC models
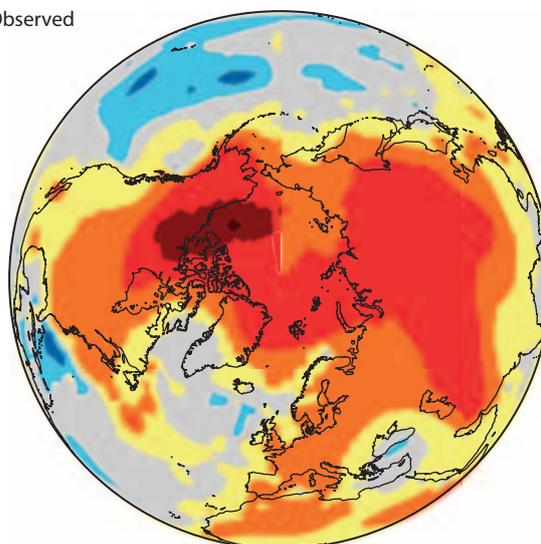
Mean temperature change, °C
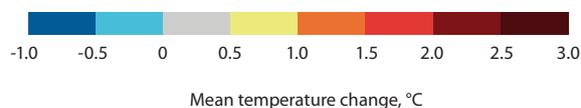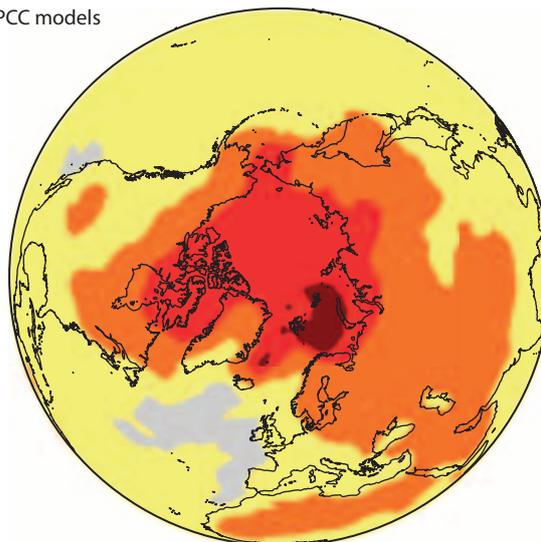
Figure 3.1. Changes in annual mean temperature over the 50-year period, 1957 to 2006, as (a) observed (NCEP/NCAR re-analysis) and (b) simulated; 'b' is a composite of fourteen IPCC model simulations in AR4 (Solomon et al., 2007). See Walsh et al. (2008) for model selection.

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 Top resolution	Ocean Resolution Z Coord., Top BC	Sea ice Dynamics, Leads (Sources)	Coupling Flux adjustments	Land 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, Germany/Korea	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)
10: FGOALS-g1.0, 2004	LASG/Institute of Atmospheric Physics, China	top = 2.2 hPa T42 (~2.8°×2.8°) L26	1.0°×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)

Model ID, Vintage	Sponsor(s), Country	Atmosphere Top resolution	Ocean Resolution Z Coord., Top BC	Sea ice Dynamics, Leads (Sources)	Coupling Flux adjustments	Land Soil, Plants, Routing (Sources)
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°×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 4°×5° L21	2°×2.5° 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 2.5° ´ 3.75° L19	12°×2° 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	0.2°×0.3° 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 (~2.8°×2.8°) L20	0.5 – 1.4°× 1.4° 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-CGCM 2.3.2, 2003	Meteorological Research Institute, Japan	top = 0.4 hPa T42 (~2.8°×2.8°) L30	0.5 – 2.0°× 2.5° 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 (~2.8°×2.8°) L26	0.5 – 0.7°× 1.1° 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 2.5°×3.8° L19	1.5°×1.5° 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 ~1.3°×1.9° L38	0.3 – 1.0°× 1.0° 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)

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

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.

among the models and across emissions scenarios. Walsh et al. (2008) 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.

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,

CCCMA

INMCM3

Composite, 14-GCM

CNRM

-3.5 -2 -1 0 1 2 3 4 5 6
Temperature change, °C
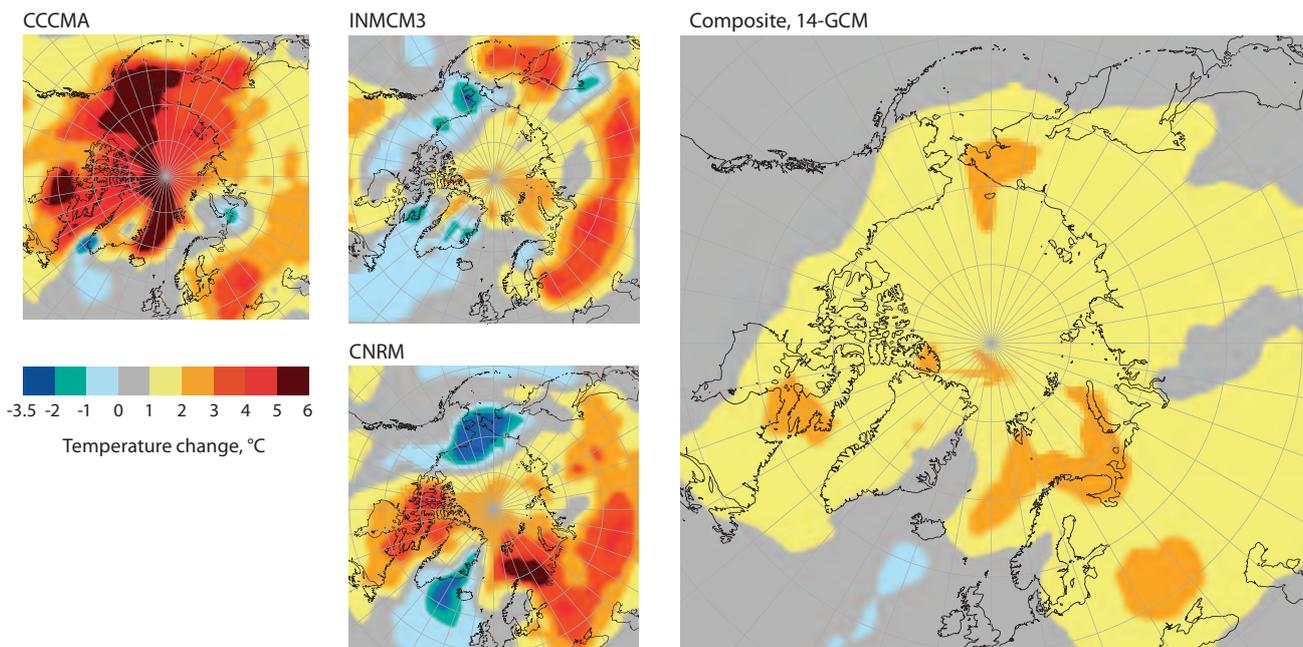
Figure 3.2. Patterns of winter temperature change over the 50-year period, 1951 to 2000, simulated by three individual models (the three small plots), and the corresponding mean pattern of fourteen different models (the large plot). All models are from the suite used in the IPCC AR4 (Solomon et al., 2007). See Walsh et al. (2008) for model selection.

Winter   Spring

Autumn   Summer

0   1   2   3   4   5   6   12

Projected temperature change, °C
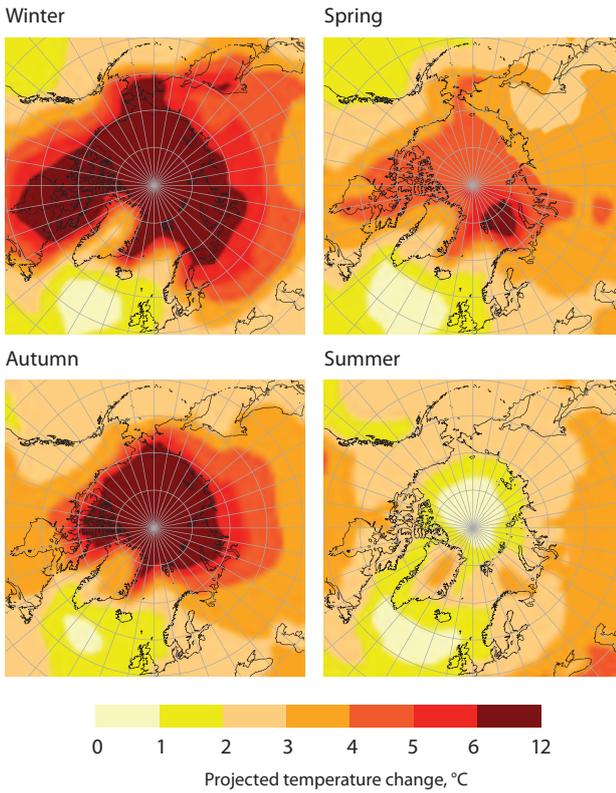
Figure 3.3. Projected changes of seasonal mean surface air temperature for the period 2070 to 2090. Changes are composited over the CMIP3 models forced by the A1B emissions scenario. Fourteen model projections form the composite. Modified from Walsh et al. (2008).

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.
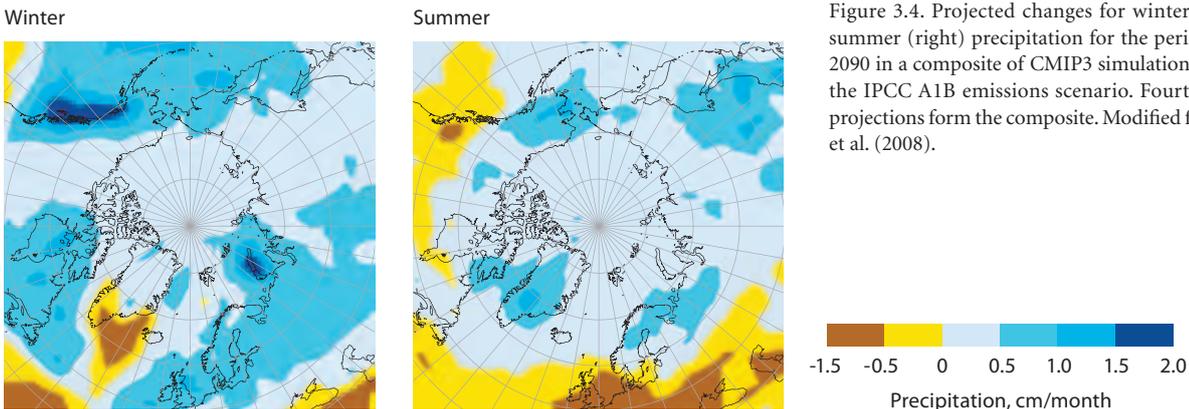
Winter   Summer

Figure 3.4. Projected changes for winter (left) and summer (right) precipitation for the period 2070 to 2090 in a composite of CMIP3 simulations forced by the IPCC A1B emissions scenario. Fourteen model projections form the composite. Modified from Walsh et al. (2008).

-1.5   -0.5   0   0.5   1.0   1.5   2.0

Precipitation, cm/month

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.

Average of the 12 monthly
RMS errors for each model, °C
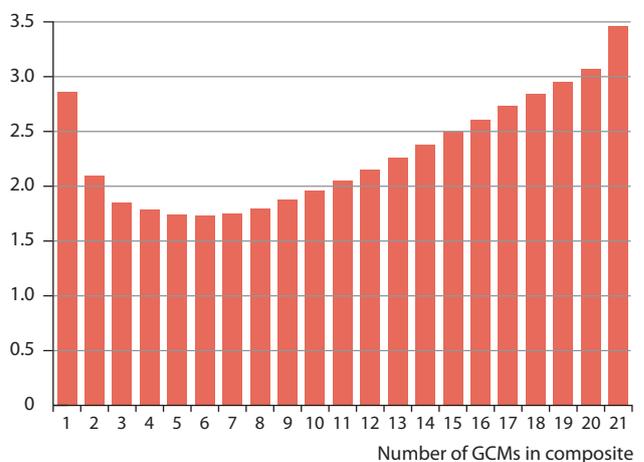
Figure 3.5. Dependence of the root mean square error (integrated over 60° to 90° N and the seasonal cycle for the period 1981 to 2000) of Arctic surface air temperature on the number of highest-ranking models included in a multi-model ensemble composite.

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., 2008). 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., 2008). 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.

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

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.

Model	Time series						Teleconnection pattern					
	AO	NAO	AMO	ENSO	PDO	PNA	AO	NAO	AMO	ENSO	PDO	PNA
BCCR												
CCSM3.0												
CGCM3.1 (T47)												
CNRM												
CSIRO-MK3												
ECHAM5												
GFDL-CM2.0												
GFDL-CM2.1												
GISS-ER												
INM-CM3												
IPSL												
MIROC-MED												
MRI												
PCM												
UKMO-HADCM3												
UKMO-HADGEM												

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 practice 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.

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., 2008		Walsh et al., 2008		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 (monthly mean)		RMS error (monthly mean)		Variance (winter)	Summer mean	Seasonal cycle	
1	BCCR-BCM2.0	Norway	16		14	● (red)						×	×	
2	CCM3	USA	7	9		● (blue)	2/2	13/13	2/3	15/4	√	√	√	2.7
3	CGCM3.1(T47)	Canada	5	6	5	● (blue)	10/10	4/5	11/8	2/4	√	√	×	3.4
4	CGCM3.1(T63)	Canada	8	8	4	● (blue)					√	√	×	3.4
5	CNRM-CM3	France	14	13	7	● (blue)	5/6	11/3	5/6	12/3	√	√	√	
6	CSIRO-Mk3.0	Australia	13	7	19	● (red)	12/13	9/10	14/13	9/14	√	×	×	3.1
7	ECHAM5/MPI	Germany	3	2	1	● (blue)	1/1	1/1	1/1	1/1	√	√	×	3.4
8	ECHO-G(MIUB)	Germany/Korea			3						√	√	√	3.2
9	FGOALS-g1.0(IAP)	China	12	20	16	● (red)	15/15	10/9	15/15	13/15	×			
10	GFDL-CM2.0	USA	10	11	9	● (red)	14/4	4/4	9/5	8/10	√	√	×	2.9
11	GFDL-CM2.1	USA	1	4	2	● (red)	5/3	2/2	3/2	4/2	√	×	×	3.4
12	GISS-AOM	USA	19	17	13	● (blue)					×	√	×	2.7
13	GISS-EH	USA	22	15	18	● (red)					×			
14	GISS-ER	USA	18	18	20	● (red)	10/12	14/14	10/10	14/11	×	×	×	
15	Ingv	Italy	11									×	×	
16	INM-CM3.0	Russia	15	14	12	● (red)	10/11	9/11	6/7	7/9	√	×	×	2.1
17	IPSL-CM4	France	17	19	21	● (red)	12/8	15/15	7/9	11/13	×	√	√	4.4
18	MIROC3.2(hires)	Japan	9	5	11	● (blue)					×	×	×	4.3
19	MIROC3.2(lowres)	Japan	12	10	8	● (blue)	3/5	5/6	4/4	3/8	×	√	√	4.0
20	MRI-CGCM2.3.2	Japan	4	12	15	● (red)	7/7	6/7	13/12	11/12	×	×	×	3.2
21	PCM	USA	21	16	17	● (red)	14/14	12/11	13/14	5/7	√			2.1
22	UKMO-HadCM3	UK	6	3	10	● (blue)	6/10	7/8	8/11	6/6	√	×	×	3.3
23	UKMO-HadGem1	UK	2	1	6	● (blue)						√	√	4.4
24	BCC-CM1	China												
25	CSIRO-MK3.5	Australia											×	

Cells with two numbers indicate the order of ranking for each model of selected variables for a 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. A blue dot indicates cases where a model fares better than the typical model with respect to the reference data, and a red dot indicates the contrary. A tick and a cross indicate where models passed or did not pass the selection criterion.

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. (2008) 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.

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 blue 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 in results 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. (2008). 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. (2008), 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
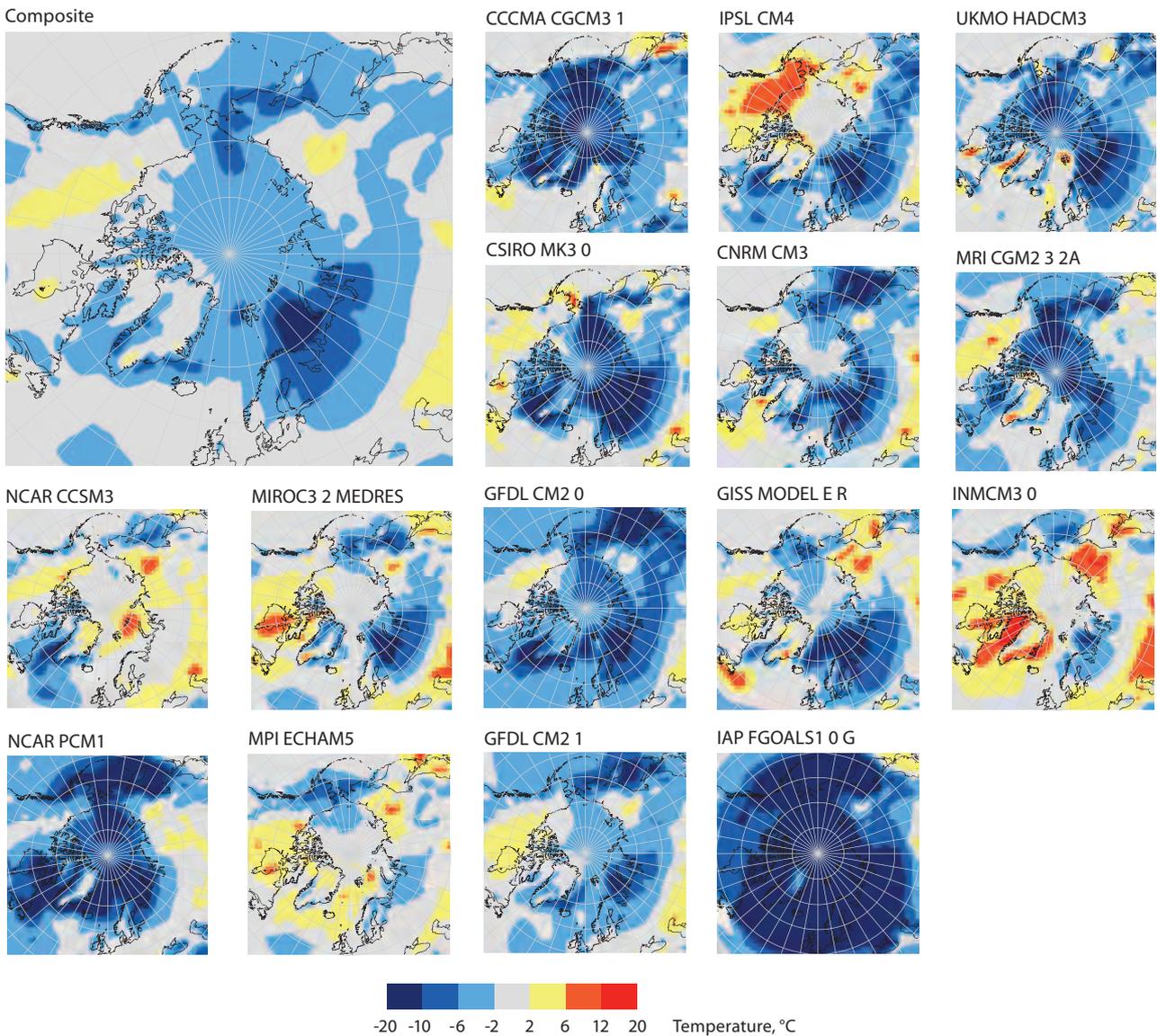
Figure 3.6. Root mean square error of January air temperatures relative to the ERA-40 re-analysis for the period 1958 to 2000. Simulations are from a subset of the 20th-century simulations run for IPCC AR4 (Solomon et al., 2007). Source: Walsh et al. (2008).

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

Barents Sea sea ice extent, 10^6 km²
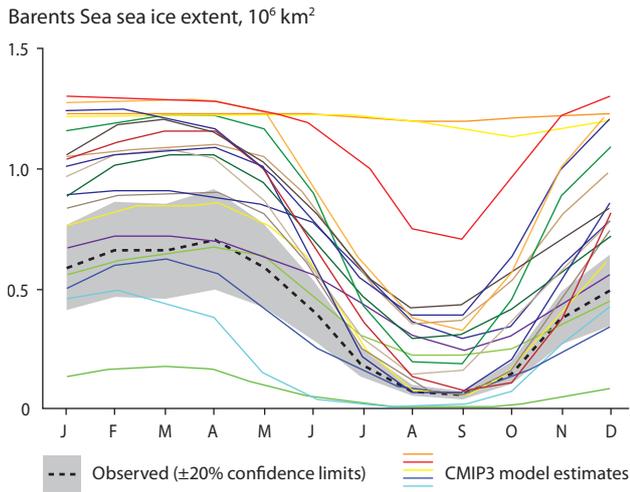
Figure 3.7. Seasonal cycle of sea ice in the Barents Sea according to various CMIP3 models. Note that most models show too much sea ice coverage. Source: modified version of figure 3 from Overland and Wang (2007).

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 (Aðalgeirsdóttir 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

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4. Changing Snow Cover and its Impacts

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Contents

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

Major advances have been made in the ability to monitor Arctic snow cover from space over the past ten years from sensors such as MODIS, AMSR-E and QuikSCAT, but a number of significant challenges remain for documenting past variability and rates of change. These include major gaps in observing networks, unrepresentative observations, and inconsistencies in analysis products. The provision of reliable precipitation information over the Arctic is an ongoing key 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 have 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 that are 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 0–15% for maximum SWE 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%).

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, will affect the capacity and operations of current and future hydroelectric developments and might solve some of the rising energy needs. 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 already 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 the general activities of Arctic residents such as transport, herding, hunting and fishing and other forms of land use.

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 limited ability to migrate as in the past, so 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 (including projections of future changes) are needed at the local scale to assist 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 11). 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). These changes are driving widespread reductions in snow cover duration, particularly in the spring period when the melt is earlier, as well as increased snow accumulation in some regions of northern Eurasia (Kitaev et al., 2005; Bulygina et al., 2009). The changes are also driving increased frequencies of 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 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), dedicated relatively little space to snow dynamics and the response of the Arctic snow cover to climate change simulated by the CMIP3 models did not receive much attention 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 synthesizes 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).

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; Callaghan et al., 2010).

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.

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 these areas 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
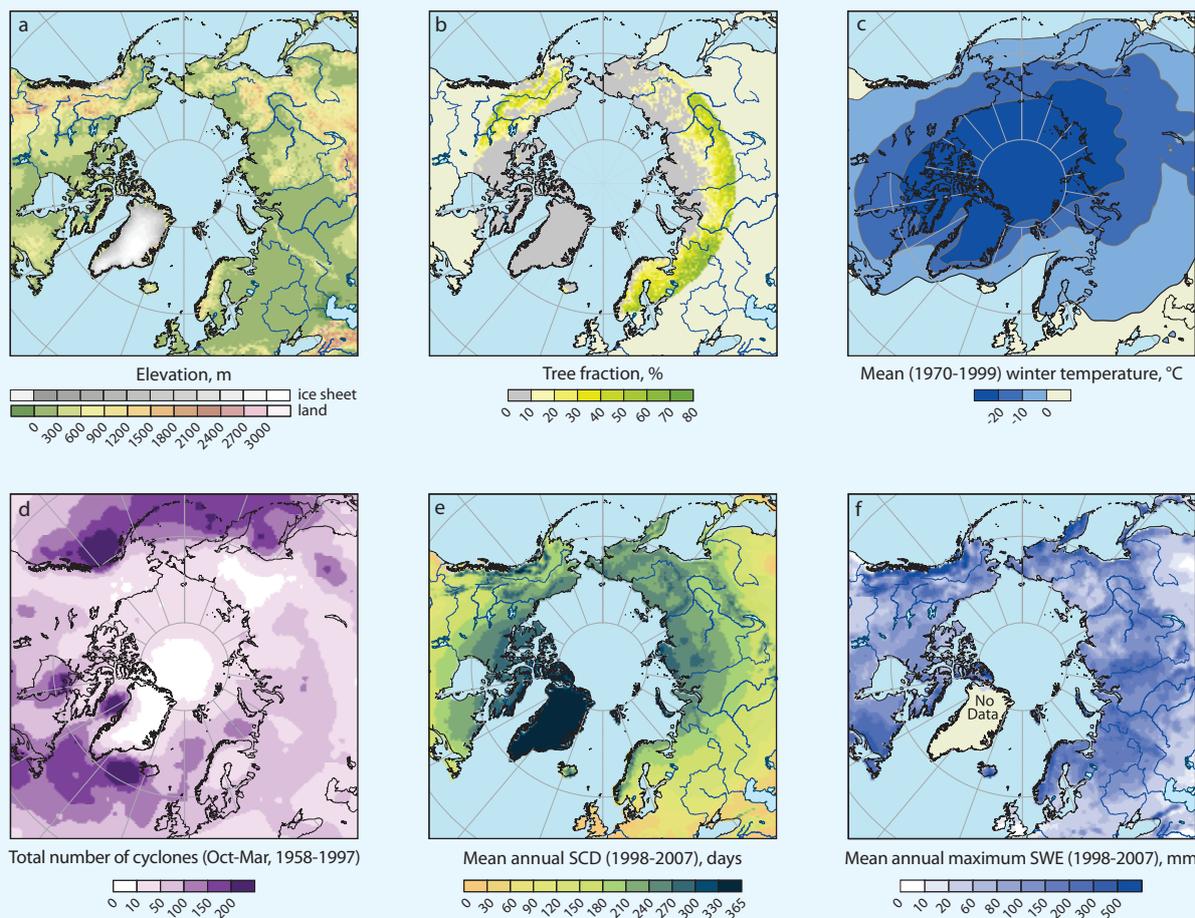
Figure 4.1. Basic characteristics of the Arctic related to snow cover: (a) elevation distribution over the Arctic from the GTOPO30 dataset, (b) tree fraction from 500-m resolution MODIS Vegetation Continuous Field product (Hansen et al., 2003), (c) 0 °C and -20 °C mean winter isotherms (November–May) for the period 1970 to 1999, from the NCEP Reanalysis, (d) October to March cyclone frequency from the NCEP Reanalysis for the period 1958 to 1997 from Key and Chan (1999), (e) mean annual snow-cover duration for the period 1998 to 2007 from the NOAA IMS 24 km product (Ramsay, 1998), (f) mean annual maximum snow-water equivalent for the period 1998 to 2007 from the Canadian Meteorological Centre snow depth analysis (Brasnett, 1999).

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/m³ 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/boreal forest), snow cover is more uniform and less dense (~200 kg/m³) 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#s). Similarly, the Sámi have a rich vocabulary, but this also relates to snowpack conditions relevant to reindeer herding practices (Ryd, 2001).

SWE, mm
Snow density, kg/m³
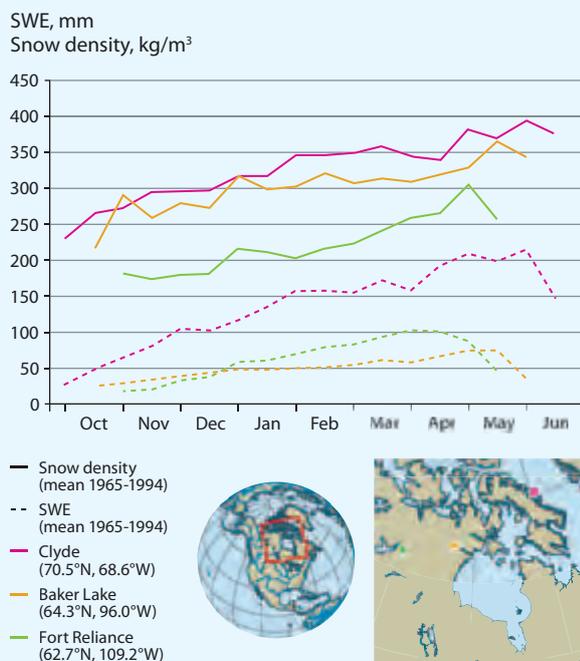
Figure 4.2. Mean seasonal evolution of snow-water equivalent and snow density for the period 1965 to 1994 for three stations along an east-west transect over the Canadian Arctic representing a typical snow season in the Arctic. Source: data from Meteorological Service of Canada (2000).

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/m² (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 SWE and SCD vary across the Arctic. The largest decreases in SWE and SCD 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 SCD is projected to decrease. Decreases by 10–20% 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 SCD and snow 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 0–15% are projected over much of the Arctic with the largest increases (15–30%) 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).

The analysis is based on monthly total precipitation data collected at the stations from the start of their operation up
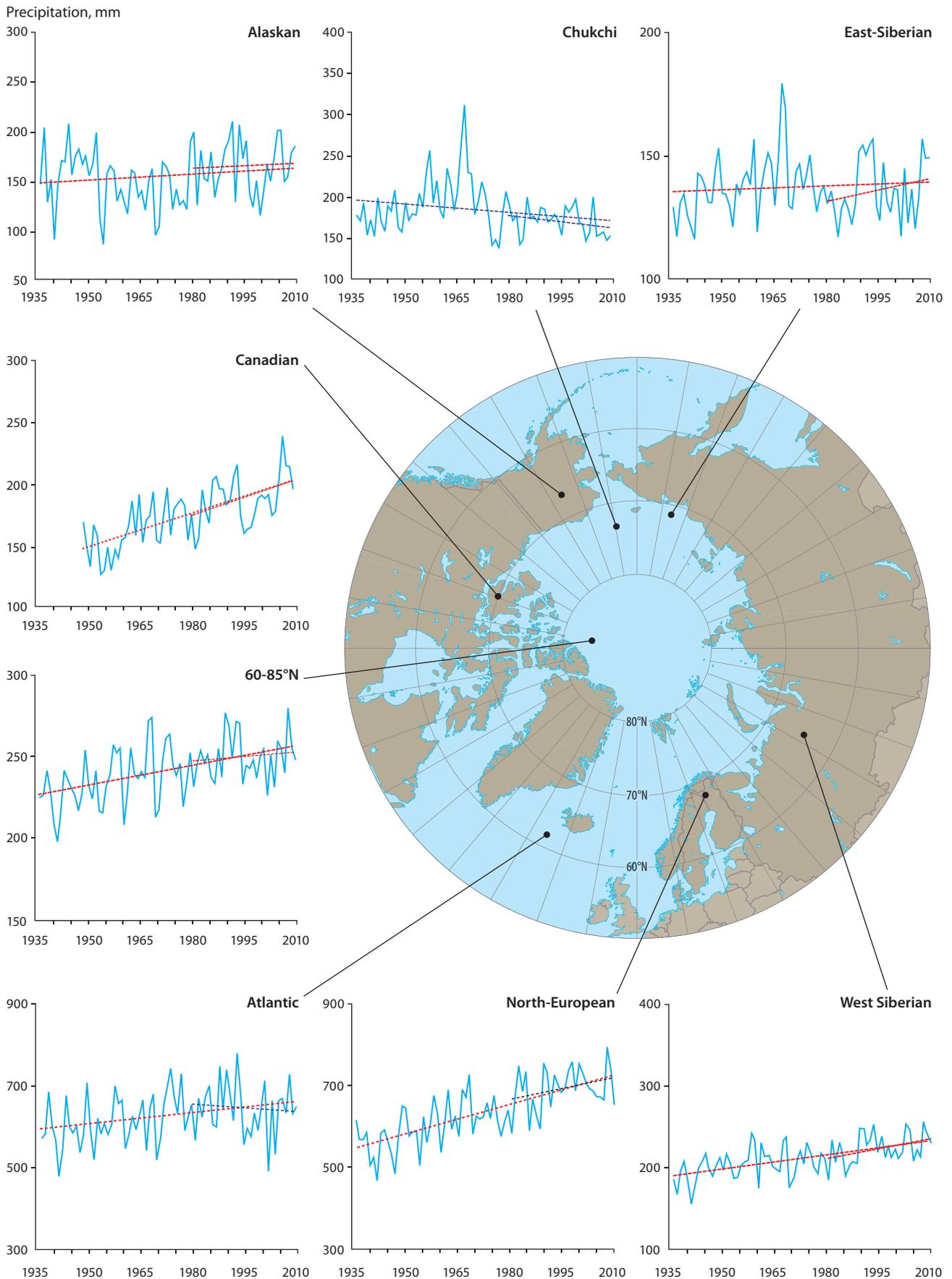
Figure 4.3. Long-term variability of total precipitation during the cold season (October to May when mean monthly temperature is below -2 °C) for inland stations in different sectors of the Arctic (see Appendix 4.1 for sector definitions) together with their long-term (1936–2009) and short-term (1980–2009) linear trends; red line denotes positive trend and dark blue line negative trend. Precipitation amounts were calculated using a method of optimal averaging (Frolov, 2010). Results for Canada used the adjusted and homogenized precipitation dataset of Mekis and Hogg (1999) updated to 2008.
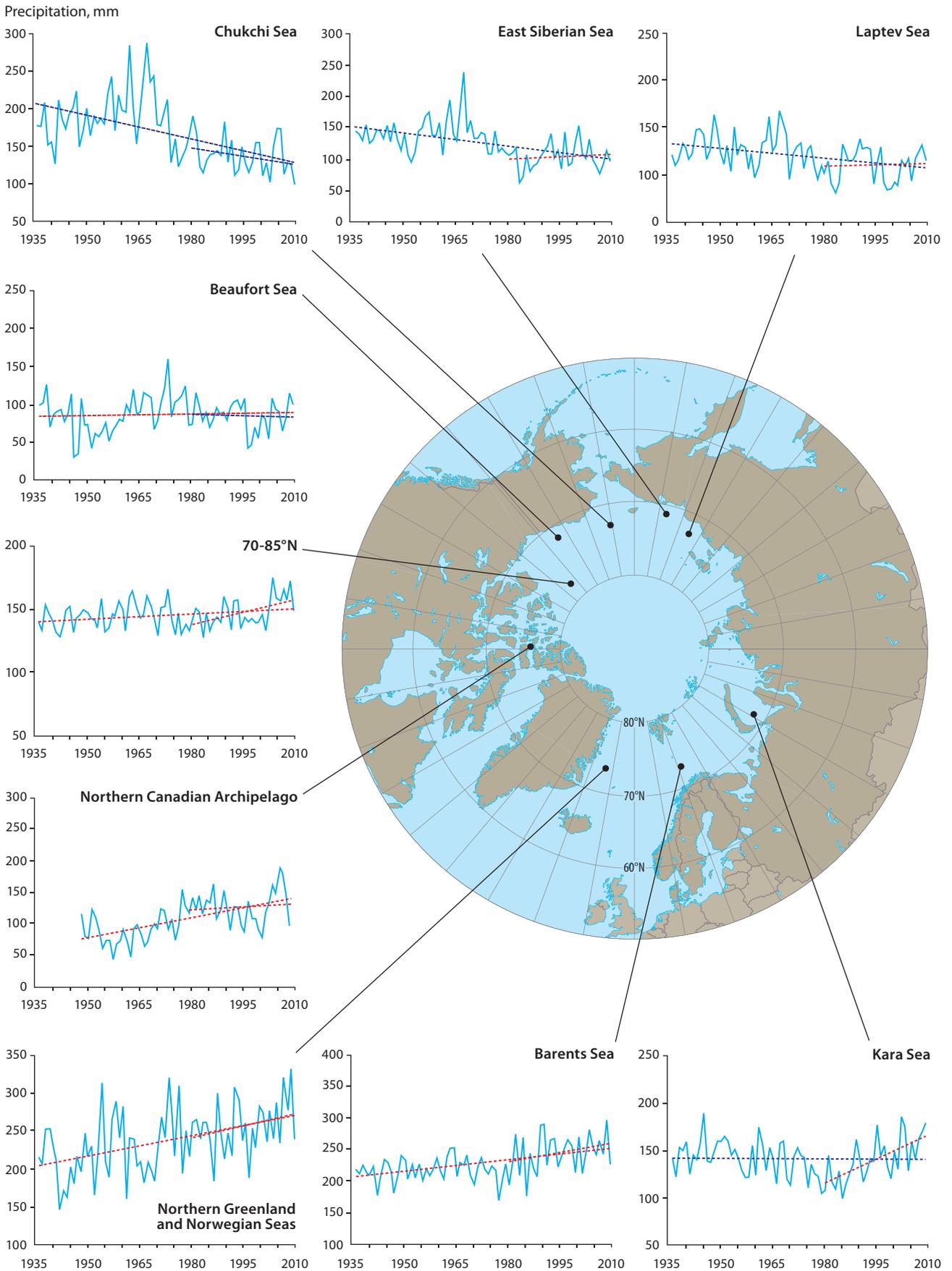
Figure 4.4. Long-term variability of total precipitation during the cold season (October to May when mean monthly temperature is below -2 °C) in the marginal Arctic seas, central Arctic basin, and northern territories of the Canadian Archipelago together with their long-term (1936 to 2009) and short-term (1980 to 2009) linear trends; red line denotes positive trend and dark blue line negative trend. Precipitation amounts over each sea were calculated using a method of arithmetic averaging of total precipitation at all coastal meteorological stations near or in the area of each sea, and at North Pole drifting stations (Frolov, 2010). Results for Canada used the adjusted and homogenized precipitation dataset of Mekis and Hogg (1999) updated to 2008.

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.

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, 2010). 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, 4.6 and 4.7) but significant decreases over the North American Arctic between 1950 and 2006 (Figure 4.8) (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 4.6). 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 4.7 and 4.8) 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. The authors concluded that paradoxically, the increase in SCD could not be associated with 'Arctic warming', which was not apparent over the period of their observation.
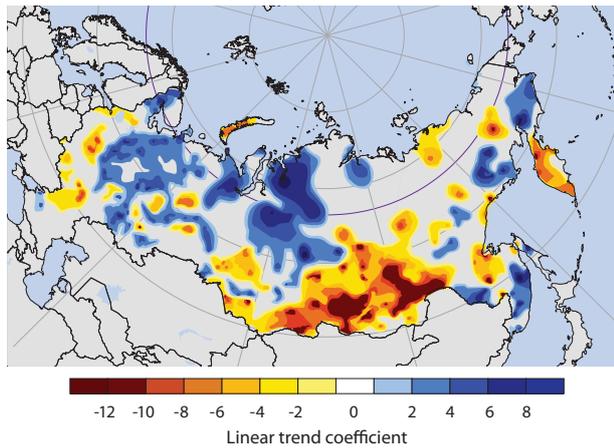
Figure 4.5. Spatial distribution of linear trend coefficients (cm per decade; 1% to 5% significance level) of maximum snow depth for permanent snow cover over the period 1966 to 2007. Source: Bulygina et al. (2009).
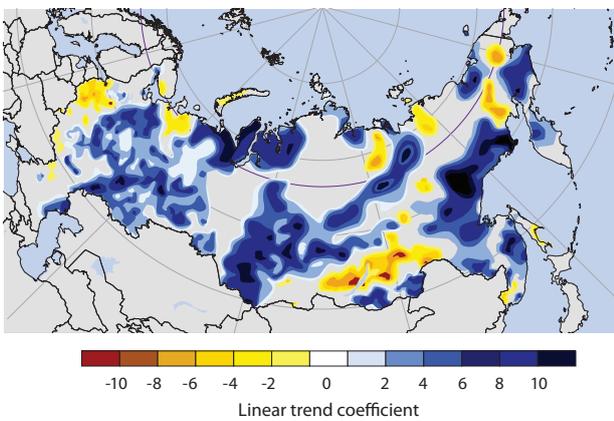
Figure 4.6. Spatial distribution of linear trend coefficients (days per decade; 1% to 5% significance level) showing the number of days with snow cover exceeding 20 cm for the period 1966 to 2007. Source: Bulygina et al. (2009).

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).

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. (2008) 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
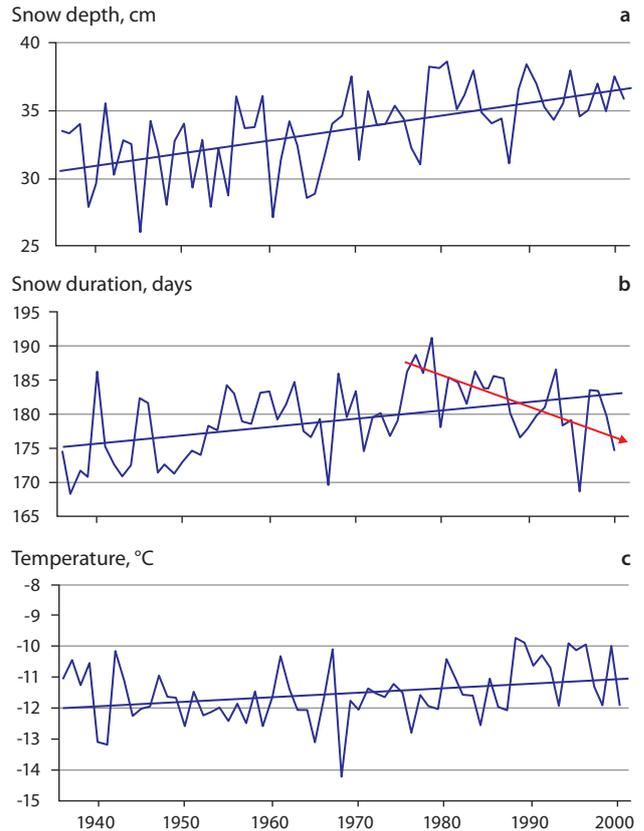
Figure 4.7. Long-term variability in northern Eurasia (1936 to 2000) of (a) mean February snow depth, (b) the number of days where 50% or more of the land surface is snow covered in winter (November to May), and (c) mean air temperature during winter (November to May). Solid lines indicate linear trends (Kitaev et al., 2005); the red line is an ad hoc indication of recent decreases in the number of days with snow cover in contrast to the long-term trend.
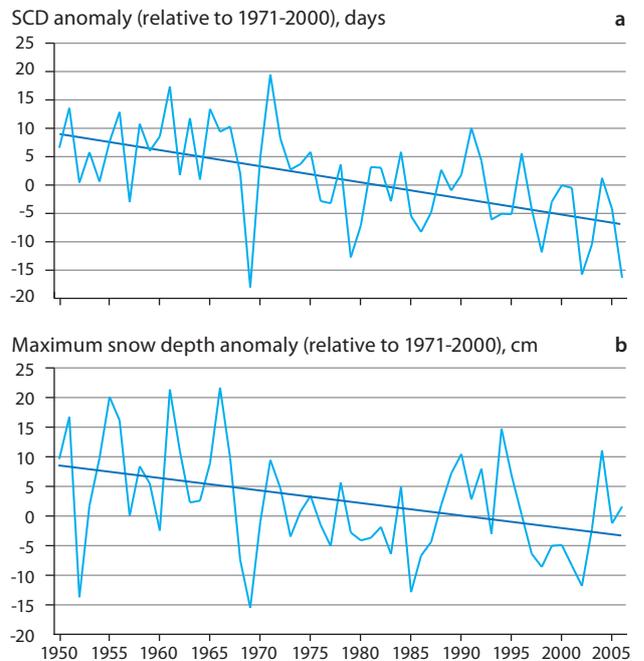
Figure 4.8. Variability and change in North American Arctic snow cover from the 1950/51 to 2006/07 snow season. (a) Annual snow cover duration. (b) Annual maximum snow depth. Values are regionally averaged anomalies from 23 stations (14 from Canada, 9 from Alaska) with anomalies calculated with respect to a 1971 to 2000 reference period. The least squares linear trend lines shown are both statistically significant (0.05 level) with slopes of -2.8 days/decade and -1.9 cm/decade. Source: updated from Atkinson et al. (2006).

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).

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*
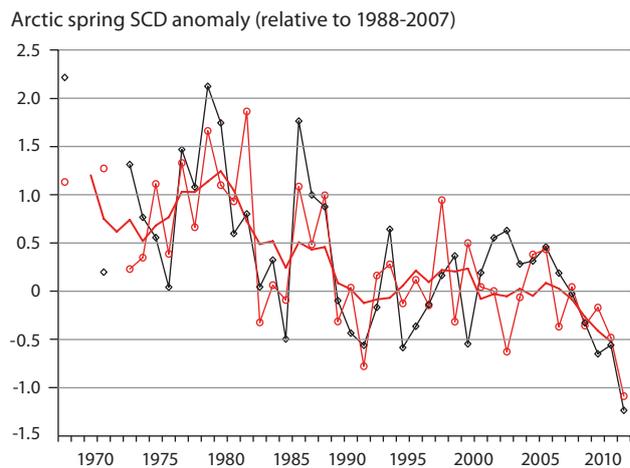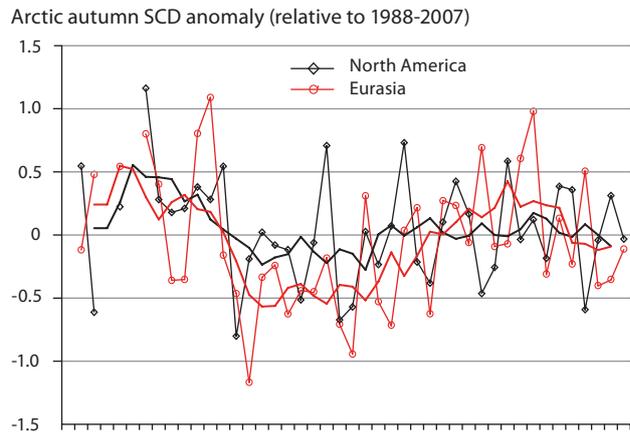
Figure 4.9. Time series of standardized snow-cover duration anomalies for the North American and Eurasian Arctic (relative to a 1988 to 2007 reference period) from the NOAA record for the (a) first half (autumn) and (b) second half (spring) of the snow season. Solid lines denote the five-year moving average. Source: Derksen et al. (2009).
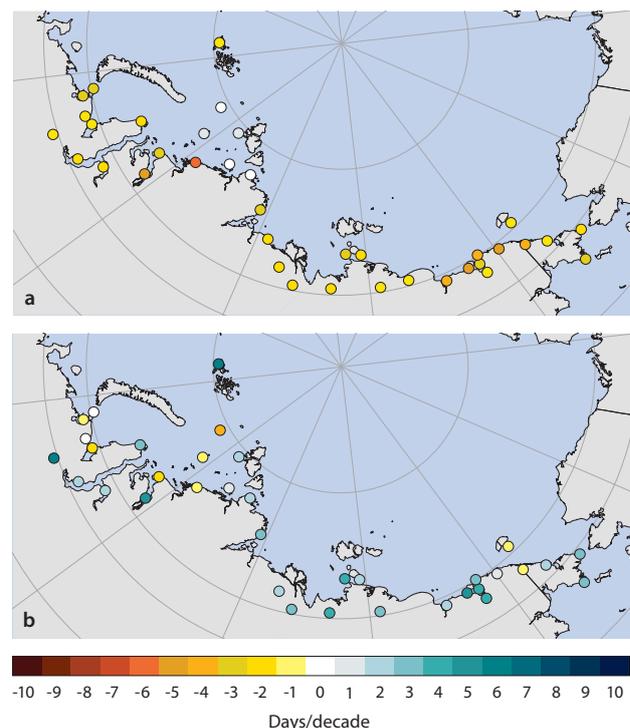
Figure 4.10. Linear trend in the (a) formation and (b) decay dates of continuous snow cover at Arctic coastal stations for the period 1973 to 2010.
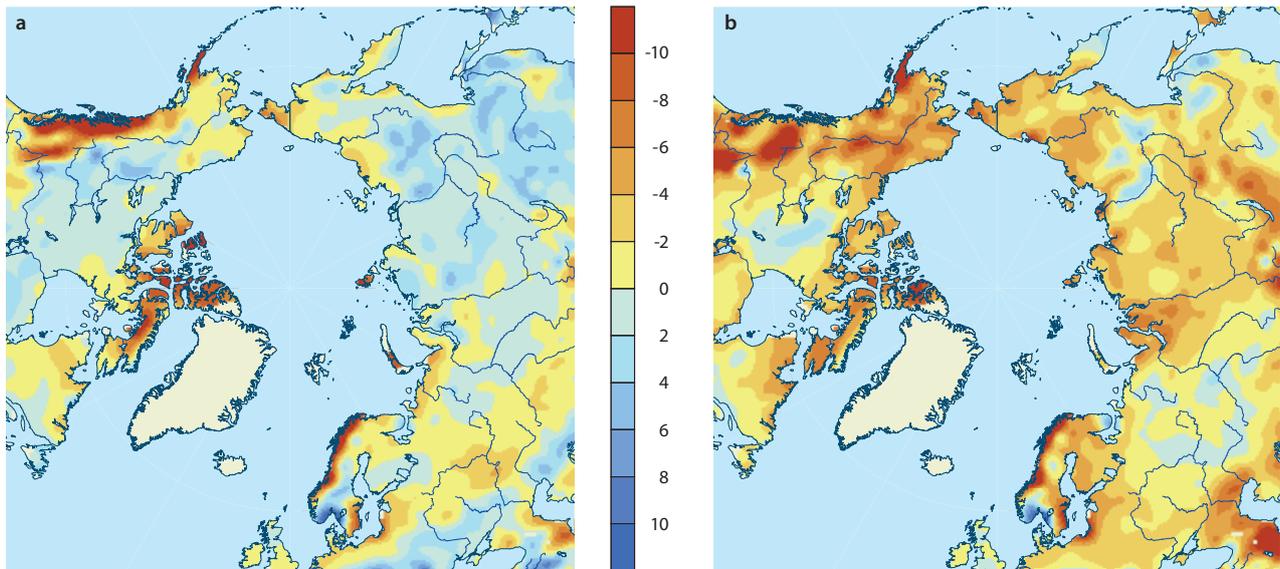
Change in SCD (1972-2008), days/decade

Figure 4.11. Trend in snow-cover duration for (a) autumn (snow-cover onset period) and (b) spring (snow-cover melt period) over the 1972/73 to 2008/09 snow seasons from the NOAA weekly dataset maintained at Rutgers University by Dr. D. Robinson. Trends were computed using the least squares method. Source: updated from UNEP (2007) by R. Brown, Climate Research Division, Environment Canada, 21 September 2009.

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., 2011). 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., 2011 and Johansson et al., 2011, 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., 2011: Johansson et al., 2011). 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
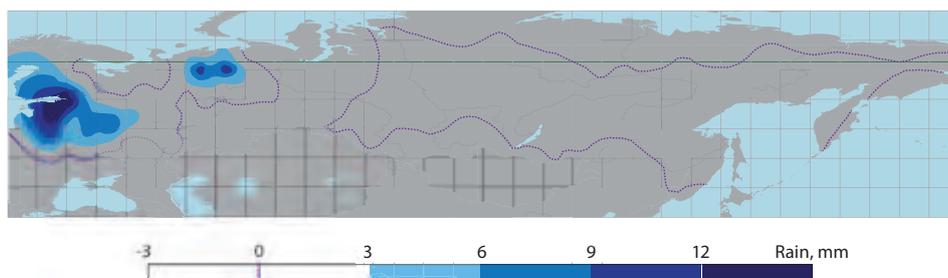
Figure 4.12. Sum of rain exceeding 4 mm per day falling on snow deeper than 3 cm, showing the change for 1989 to 2006 relative to 1951 to 1980. Source: Shmakin (2010).

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. In contrast, a 50-year record of snow stratigraphy from northern Sweden shows a significant increase in hard snow/ice layers in the snow pack, particular at the ground surface (Johansson et al., 2011).

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 (2006).

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 projected 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

Figure 4.13. Percentage of 14 CMIP3 climate models showing (a) significant decreases, (b) no increases and (c) significant increases in annual maximum snow-water equivalent for the 2080 time horizon (2070–2099 minus 1970–1999) based on the IPCC A2 emissions scenario. Source: Brown and Mote (2009).

Percentage of models with significant decrease in SWEmax

Percentage of models with no significant change in SWEmax

Percentage of models with significant increase in SWEmax

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, 2006), 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
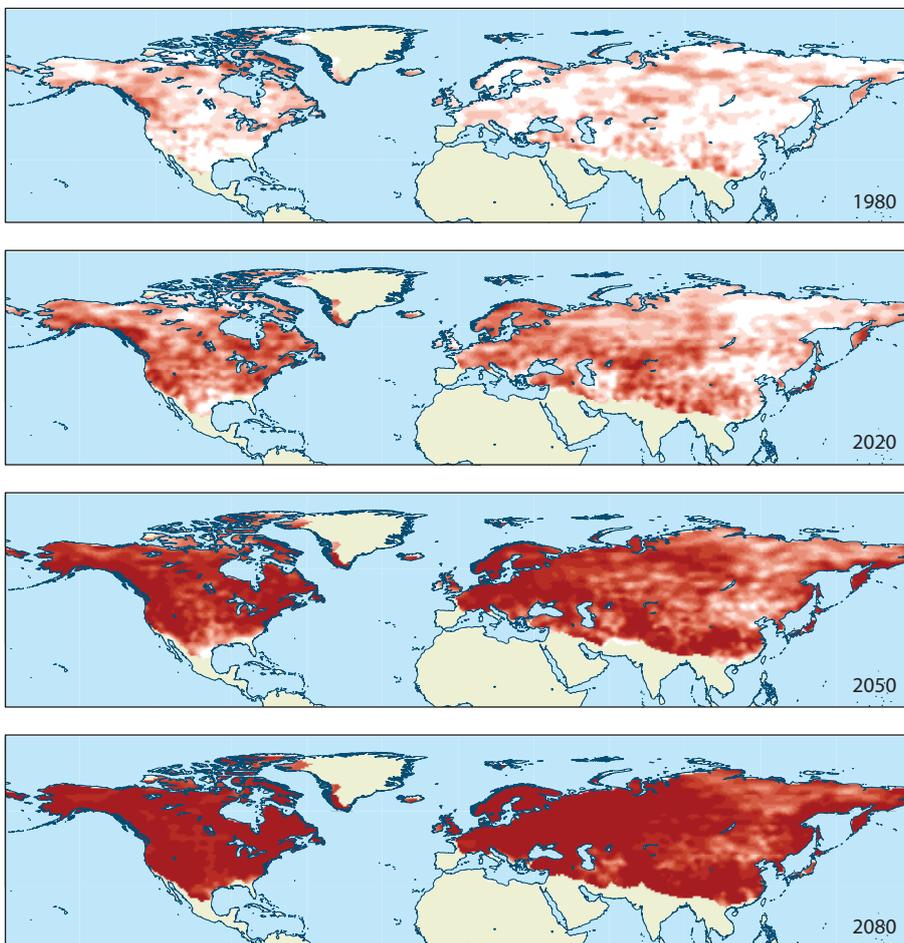
Figure 4.14. Evolution of the climate-warming signal in annual snow-cover duration from eight CMIP3 climate models (showing the percentage of models that show statistically significant local decreases in annual snow-cover duration). The top panel results are for 1970–1999 versus 1900–1929 from 20C3M simulations, while the three lower panels are computed from 30-year averages centered on the indicated date from the IPCC A2 emissions scenario with respect to a 1970–1999 reference. There is no model consensus for significant increases in SCD. Source: Brown and Mote (2009).

Percentage of models with significant decrease in SCD

10 20 30 40 50 60 70 80 90
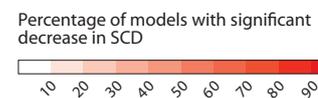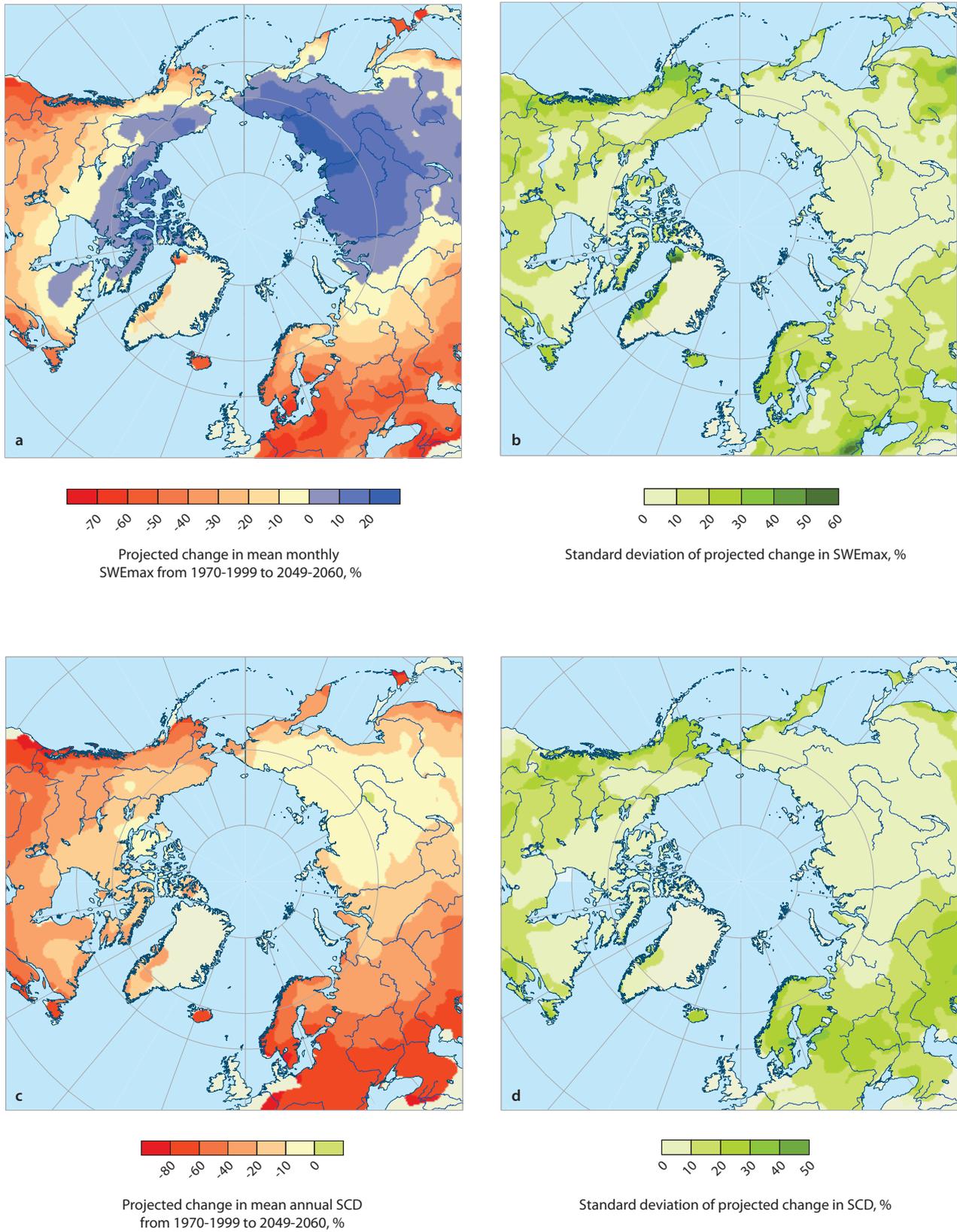
Figure 4.15. Projected changes in maximum monthly snow-water equivalent and annual snow-cover duration between 1970–1999 and 2049–2060 for six general circulation models using the IPCC A2 emissions scenario. (a) Projected percentage change in maximum monthly snow-water equivalent with the standard deviation of the six projections shown in (b). (c) Projected percentage change in mean annual snow-cover duration with the standard deviation of the six projections shown in (d). Seasonal snow-cover information over the Greenland Ice Sheet is not available from the climate models. Source: computed from climate model projections presented by Brown and Mote (2009).

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 air temperature 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). 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
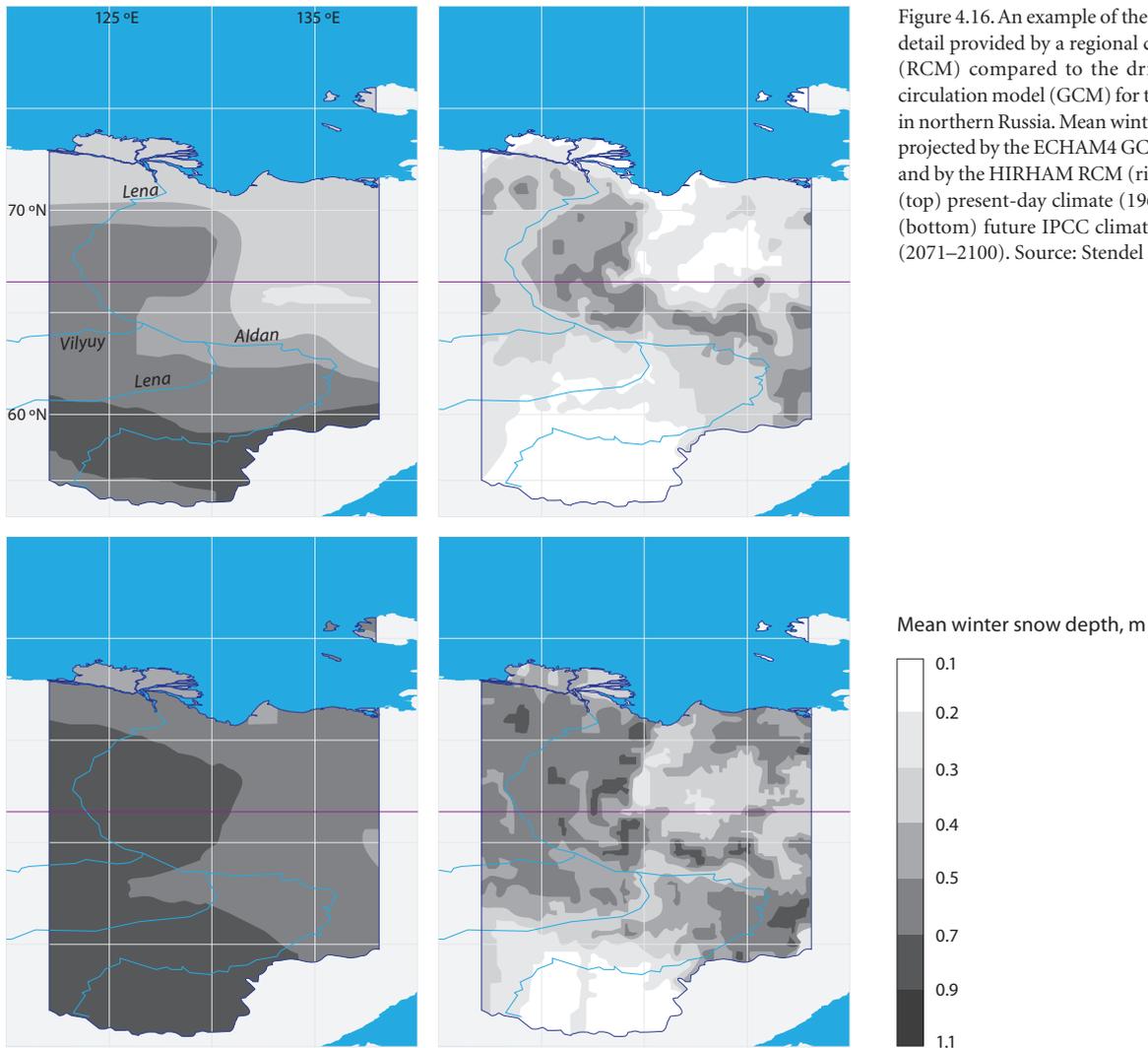
Figure 4.16. An example of the higher spatial detail provided by a regional climate model (RCM) compared to the driving general circulation model (GCM) for the Lena Basin in northern Russia. Mean winter snow depth projected by the ECHAM4 GCM (left panel) and by the HIRHAM RCM (right panel) for (top) present-day climate (1961–1990) and (bottom) future IPCC climate scenario B2 (2071–2100). Source: Stendel et al. (2007).

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 SCD 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 SCD 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 in some areas are already increasing the moisture stress of northern coniferous forests in summer.

- Winter thaws and rain-on-snow events are already damaging vegetation and impacting Arctic grazing mammal population. Projected increase in frequency of these events 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:
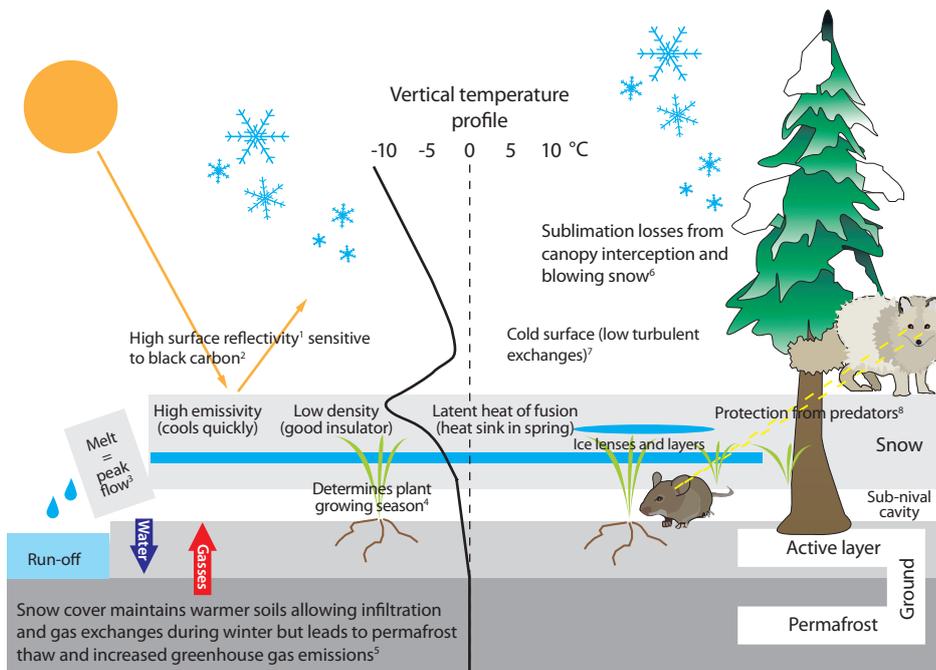
Figure 4.17. Schematic illustration outlining the importance of snow on climate, hydrology, and ecosystems. The solid black line represents an idealized ground-snow-atmosphere temperature profile to highlight the strong temperature gradients occurring near the snow surface. Source: R. Brown and T. V. Callaghan, March 2010. Superscript numbers refer to sections in the text that describe the processes as follows: [1](4.2.2), [2](4.3.1.3.2), [3](4.4.2.2), [4](4.4.3.1), [5](4.4.3.3), [6](4.4.3.1), [7](4.2.2), [8](4.4.3.2).

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 translate 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
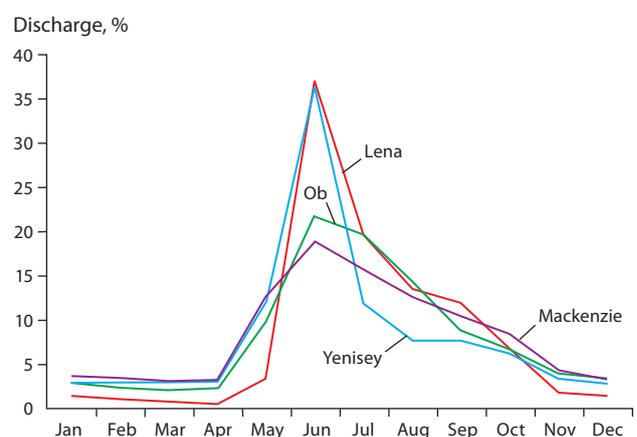
Figure 4.18. The seasonal cycle of discharge for the four largest rivers flowing into the Arctic Ocean. Source: Shiklomanov et al. (2007).

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 (see Section 4.3.1.3.1) further influence winter snow accumulation. The polar sea ice is retreating at a dramatic 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 break-up 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
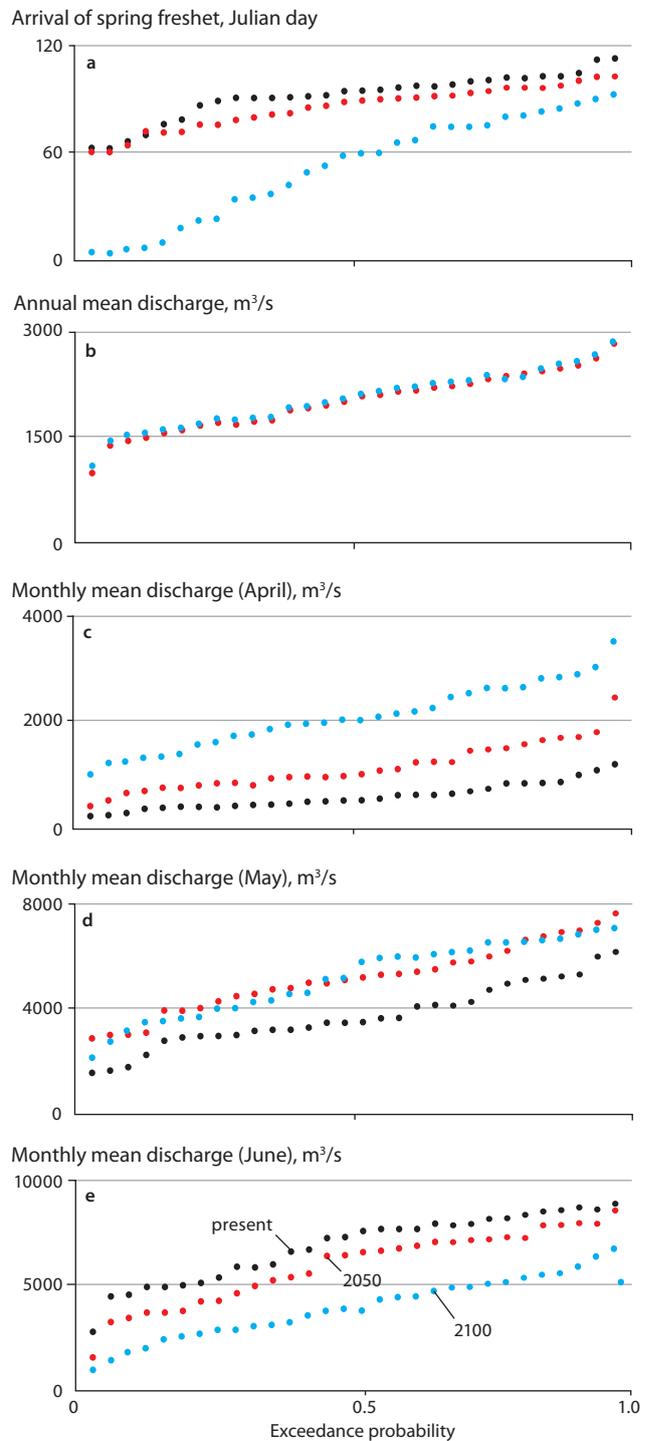
Figure 4.19. Exceedance probability (the probability that a certain value is going to be exceeded) of several streamflow characteristics of the Liard River (Yukon, Canada) for the present climate, under the IPCC B2 emissions scenario for 2050, and 2100. Showing the (a) date of arrival of spring freshet, (b) annual flow, and monthly flows for (c) April, (d) May, and (e) June. Source: Woo et al. (2008).

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.

Figure 4.20. Intercepted snow as shown in Kuusamo, northern Finland, can remain in a forest canopy for some time. Photo: Licensed under a Creative Commons CC BY SA License; Author: Muu-karhu.

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 (see Section 4.3.1.3.1). 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 that reduced the 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 et al., 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
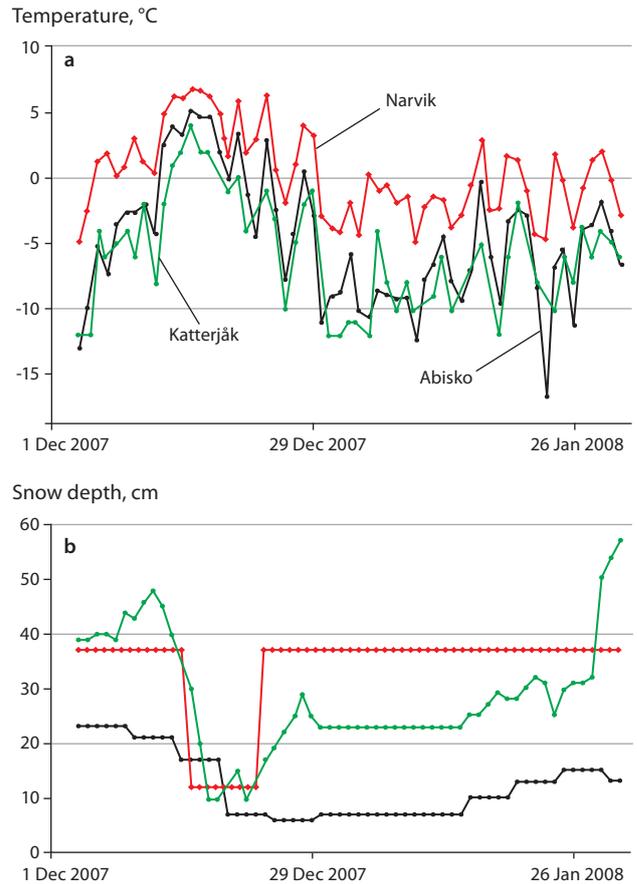
Figure 4.21. A mid-winter warming event in December 2007 and its impact on vegetation over 1400 km^2 in sub-Arctic Norway and Sweden. Air temperatures (a) increased at three sites along a transect from Narvik, Norway, to Abisko, Sweden, (68° N) for a few days in December and snow melted (b) before returning. The few snow-free days resulted in the death of some dwarf shrubs in open, alpine habitats (photo left) and coastal pine forests (photo right). Snow remaining in depressions protected vegetation (left insert); whereas unprotected vegetation died (right insert). Source: Bokhorst et al. (2009).

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
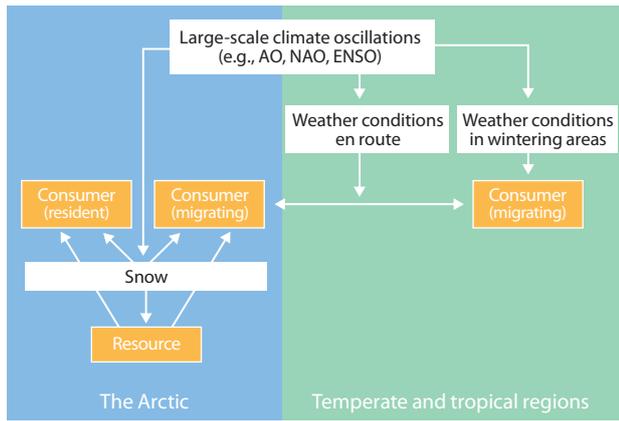
Figure 4.22. Schematic illustration contrasting the potential interactions between snow and resident Arctic animal species and between snow and migrating animal species that breed in the Arctic but overwinter in temperate or tropical regions. Changes in local weather conditions in wintering areas, en route, and in the Arctic breeding grounds have been coupled to changes in large-scale teleconnection patterns such as the Arctic Oscillation (AO), the North Atlantic Oscillation (NAO), and the El Niño Southern Oscillation (ENSO). Source: Forchhammer and Post (2004). This indicates that climate change effects embrace considerable geographical ranges. However, the consequences of land-use change and sea level rise have not been included in the scheme and could have further effects.

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 moschatu*s) (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
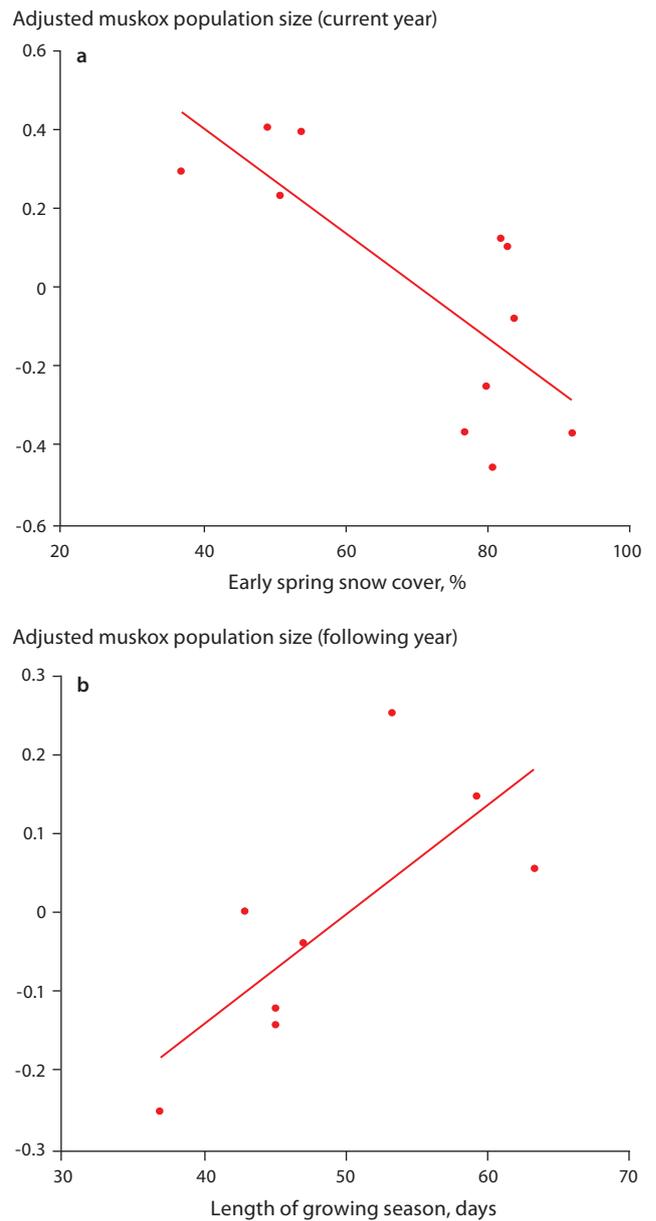
Figure 4.23. Contrasting effects of snow cover on the musk ox population at Zackenberg, northeast Greenland: (a) direct negative effect of early spring snow cover (10 June) on current-year population, and (b) delayed positive effect of increased length of the growing season on the population the following year. The length of the growing season is defined as the period with positive ecosystem assimilation and is closely related to the date of snow melt (r = -0.84, n = 8, p < 0.05). Annual population numbers of musk ox are given as relative values adjusted for (a) density dependence and (b) density dependence and spring snow cover. Data from Forchhammer et al. (2008).

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 hardness 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 mid-winter thaw on 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 high 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
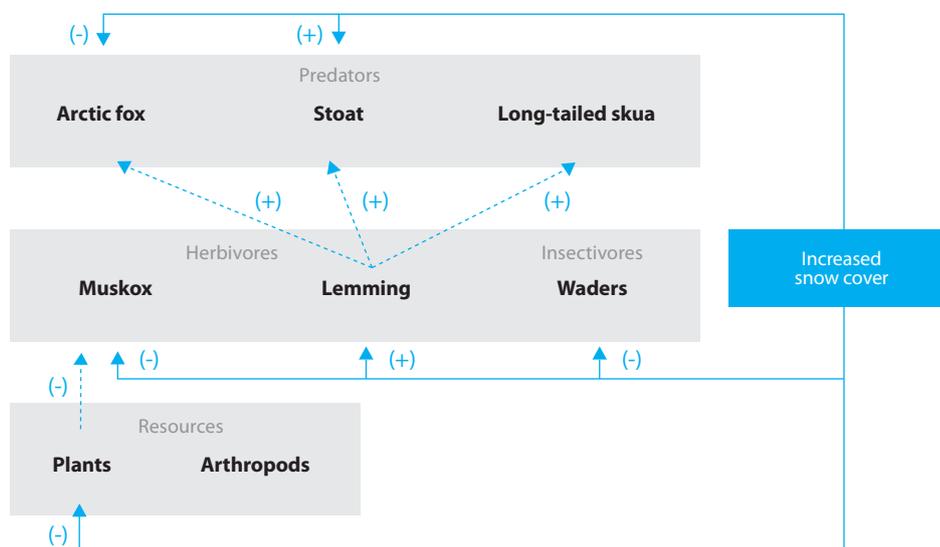
Figure 4.24. Schematic overview of the consequences of increased snow cover on the terrestrial vertebrate community at Zackenberg, northeastern Greenland. Arrows indicate potential direct (solid arrows) and indirect (dashed arrows) effects. Source: Forchhammer et al. (2008).

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 et al., 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).

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, contributing to 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.

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

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

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

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 the number of 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), an 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 et al., 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 et al., 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/m², 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 m³ of timber (Bakke 1989).

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 physically-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 possible 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 snowscapes, mountain 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.

Figure 4.25. Images based on a rare recording of a slush torrent in June 1996 from sub-Arctic Sweden. The first three images show the result of increased saturation of the snowpack and then release of a large amount of water. The fourth image shows the valley in spring with a fan formed by such short-lived events over millennia. Photo: Gude and Scheerer (1995).

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). 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., 2011). 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., 2011). 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; 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) (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 and Ray, 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 et al., 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 et al., 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 melt, 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 melt 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 et al., 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

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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., 2011). 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).
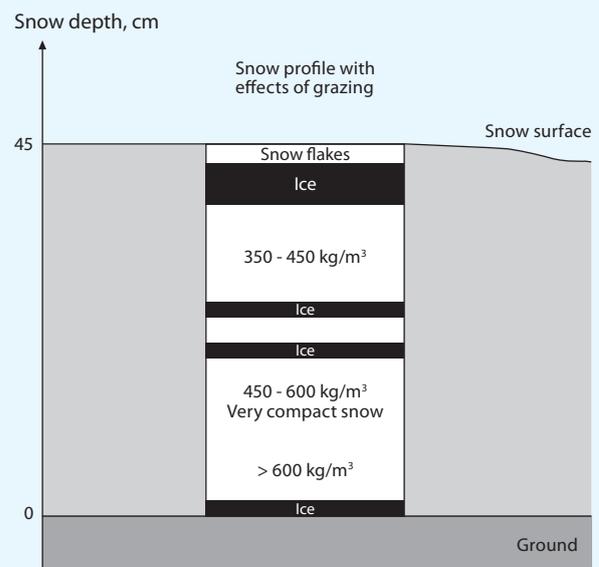
Figure 4.26. Snow-reindeer interactions: (upper left) reindeer have difficulty digging through deep snow to find food (UNEP, 2007); (upper right) Sámi reindeer herders probe the snowpack to look for problematic layers of compacted snow and ice (Photo: Terry V. Callaghan, Royal Swedish Academy of Sciences); and (lower) examples of snow stratigraphy (Riseth et al., 2011).

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 levels create archipelagos.

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 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 melt 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) 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., 2011).

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.
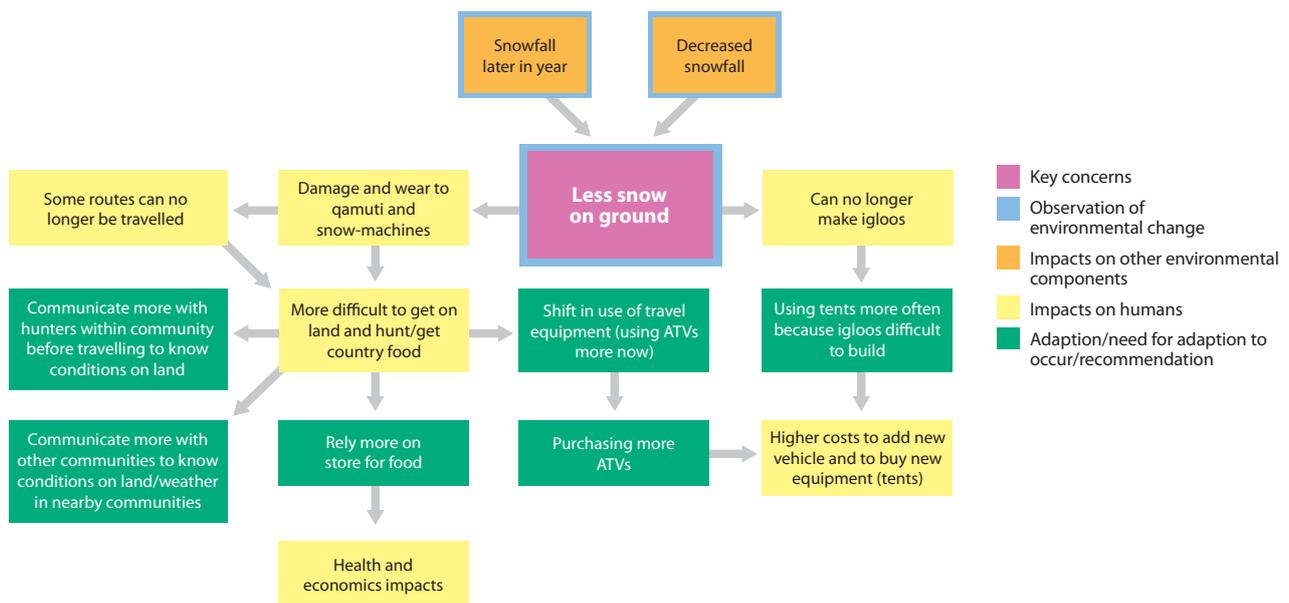
Figure 4.27. An observation, impact, and adaptation diagram for a reduced snow-cover season at Arctic Bay, Nunavut. Source: Nickels et al. (2005).

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 could 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 (within networks such as INTERACT www.eu-interact.org), 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 process (and hence under-represented topic in this assessment) 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) be carried out for applied snow-related information, such as winter thaw events, snow season start and end date, maximum SWE, date of maximum SWE, snow melt run-off timing and amount. 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 in some areas of snow clearance from roads, airstrips, etc. Current High Arctic areas 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. 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.

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.

Acknowledgments

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
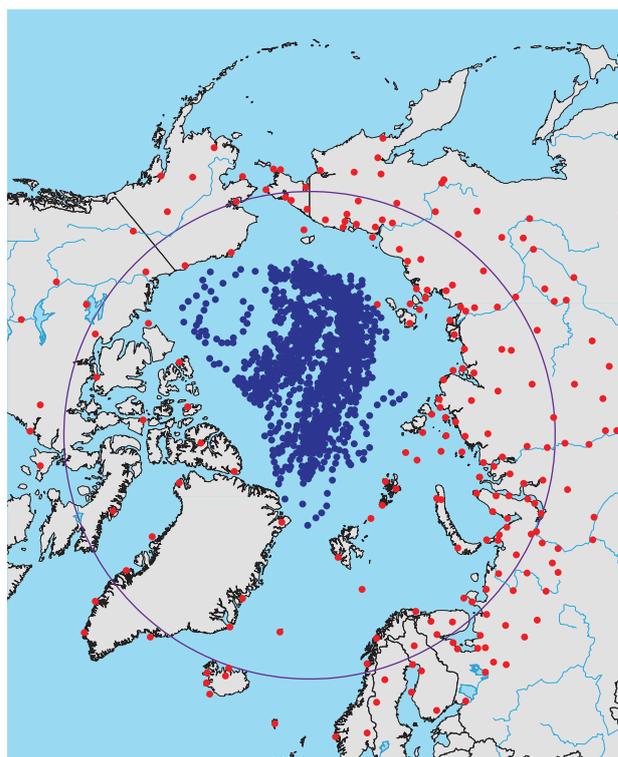
Figure A2. Long-term meteorological stations in the Arctic (red dots) and mean monthly location of North Pole drifting stations (blue dots) during the period 1937 to 2008. Source: Frolov (2010).

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

Figure A1. Spatial distribution of daily snow depth observations in the Global Summary of the Day dataset (NCDC) for March 1991–2000. Source: Ross Brown, Climate Research Division, Science and Technology Branch, Environment Canada, March 2010.

al. (2007) and Brown and Armstrong (2008), and a summary of currently available Arctic snow data sources is provided in Table A2. 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).

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)
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).
PolarView, ESA	Snow cover extent mapping over Eurasian land areas of the Arctic. (www.polarview.org/services/sm.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)

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., 2011) 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 data 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. (2009) 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., 2011).

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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Contents

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 57% 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 since the late 1970s, 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 that by the end of the 21st century the upper 2 to 3 m of permafrost will thaw over 16-20% of the area underlain by permafrost in Canada. 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 km² (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 ºC 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 to 1850 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
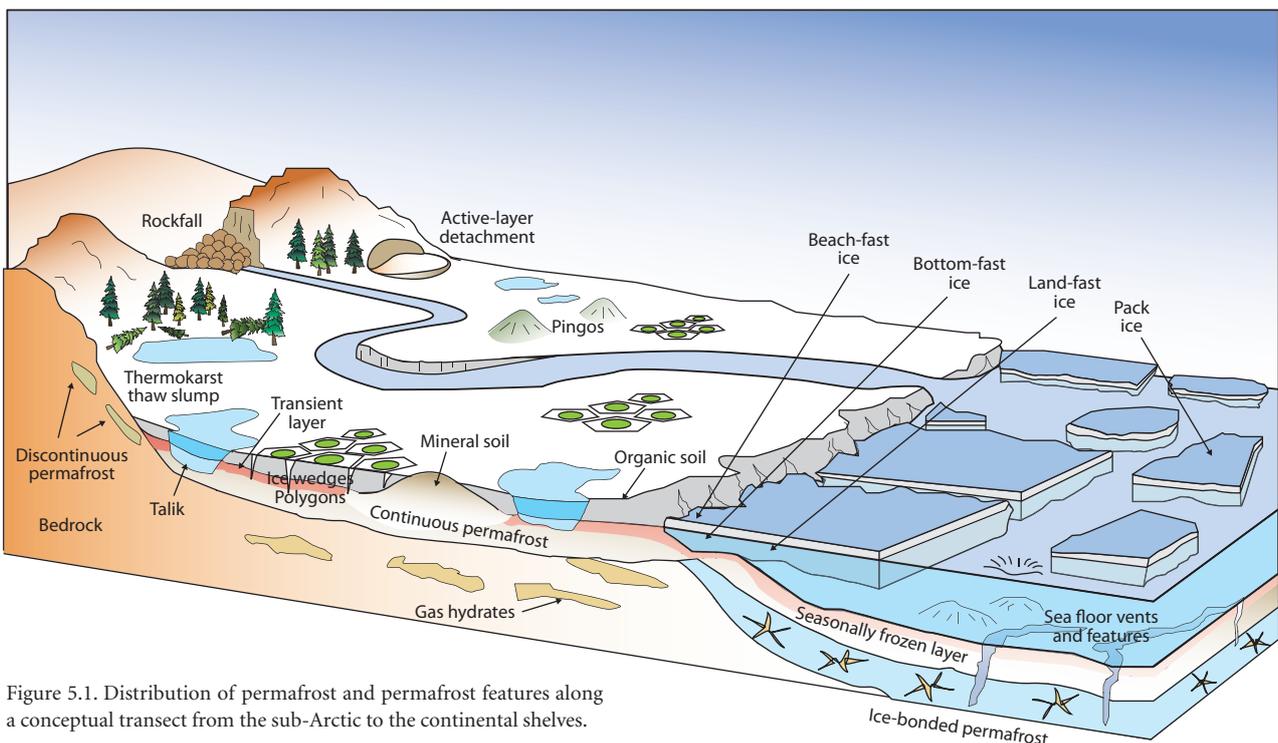
Figure 5.1. Distribution of permafrost and permafrost features along a conceptual transect from the sub-Arctic to the continental shelves.

Extent of permafrost, % of area

Continuous (90-100%)
- Continuous[1]
- Continuous[2]

Discontinuous (50-90%)
- Discontinuous[1]
- Discontinuous[2]

Sporadic (10-50%)
- Sporadic[1]
- Sporadic[2]

Isolated patches (0-10%)
- Islolated[1]
- Isolated[2]

- Subsea permafrost
- Arctic glaciers and ice sheets
- Deep relict permafrost (where known)

[1]Lowlands, highlands and intra-intermontane depressions characterized by thick overburden cover (>5-10 m)

[2]Mountains, highlands, ridges and plateaus characterized by thin overburden cover (<5-10m) and exposed bedrock

Figure 5.2. Extent and type of permafrost in the Northern Hemisphere. Source: International Permafrost Association, Brown et al. (1998).

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 provides a synthesis of 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 since the late 1970s, 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:

Last Interglacial optimum (125 ky BP)
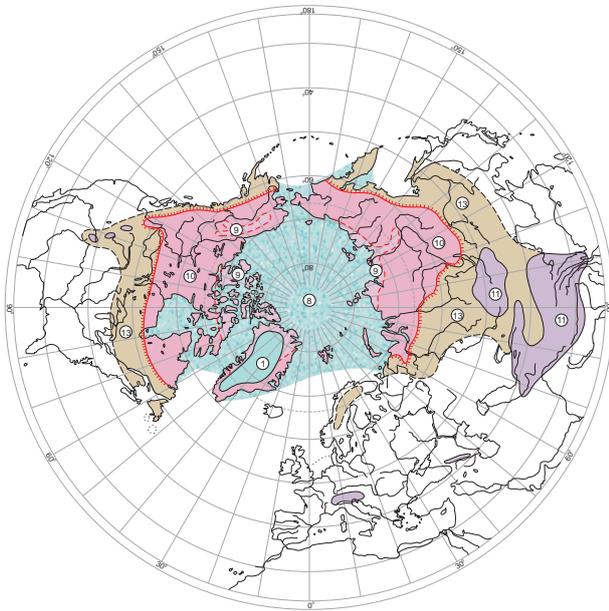
Last glacial maximum (20 - 18 ky BP)

Holocene (6 - 5.5 ky BP)

1	Ice sheets		10	Discontinuous and sporadic permafrost
2	Mountain glacial complexes		11	Alpine permafrost
3	Areas of alpine glaciers		12	Areas of permafrost with local glaciers (on Arctic islands)
	Ice divides		13	Areas of deep seasonal freezing
	Ice lobes		400	Isolines of permafrost thickness
	Ice flow direction		-3-	Geoisotherms of permafrost
7	Perennial sea ice			Boundary of continuous permafrost
8	Seasonal sea ice			Southern limit of permafrost
9	Continuous permafrost; thickness: ‡ – reconstructed · – non-reconstructed			Ancient coastline

Figure 5.3. Permafrost distribution during the last interglacial period (125 ky BP), the last glacial maximum (20 to 18 ky BP), and during the Holocene climatic optimum (6 to 5.5 ky BP) based on paleo-reconstructions by Velichko and Faustova (2009), and Velichko and Nechaev (2009).

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, 2009). 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

Figure 5.4. Present-day distribution of permafrost of different ages in Russia. Source: after Lisitsyna and Romanovskii (1998).

■ Submarine permafrost

■ Pre-Holocene permafrost

■ Holocene permafrost

■ Relict permafrost
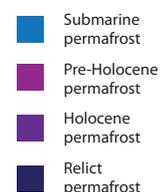
causing localized thawing under those lakes that were sufficiently deep (MacDonald et al., 2006; Walter et al., 2007a). The rate of thawing both in the vertical and lateral direction was several centimetres to several tens of centimetres per year. Several thousand years after the Holocene optimum, permafrost 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
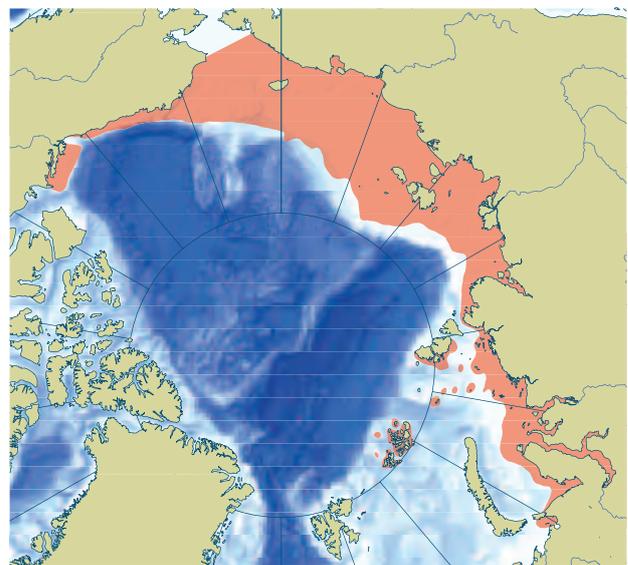
Figure 5.5. Extent of subsea permafrost distribution (orange), superimposed on bathymetry. Map based on Brown et al. (1998). Note: see Figure 5.2 for details of terrestrial permafrost.
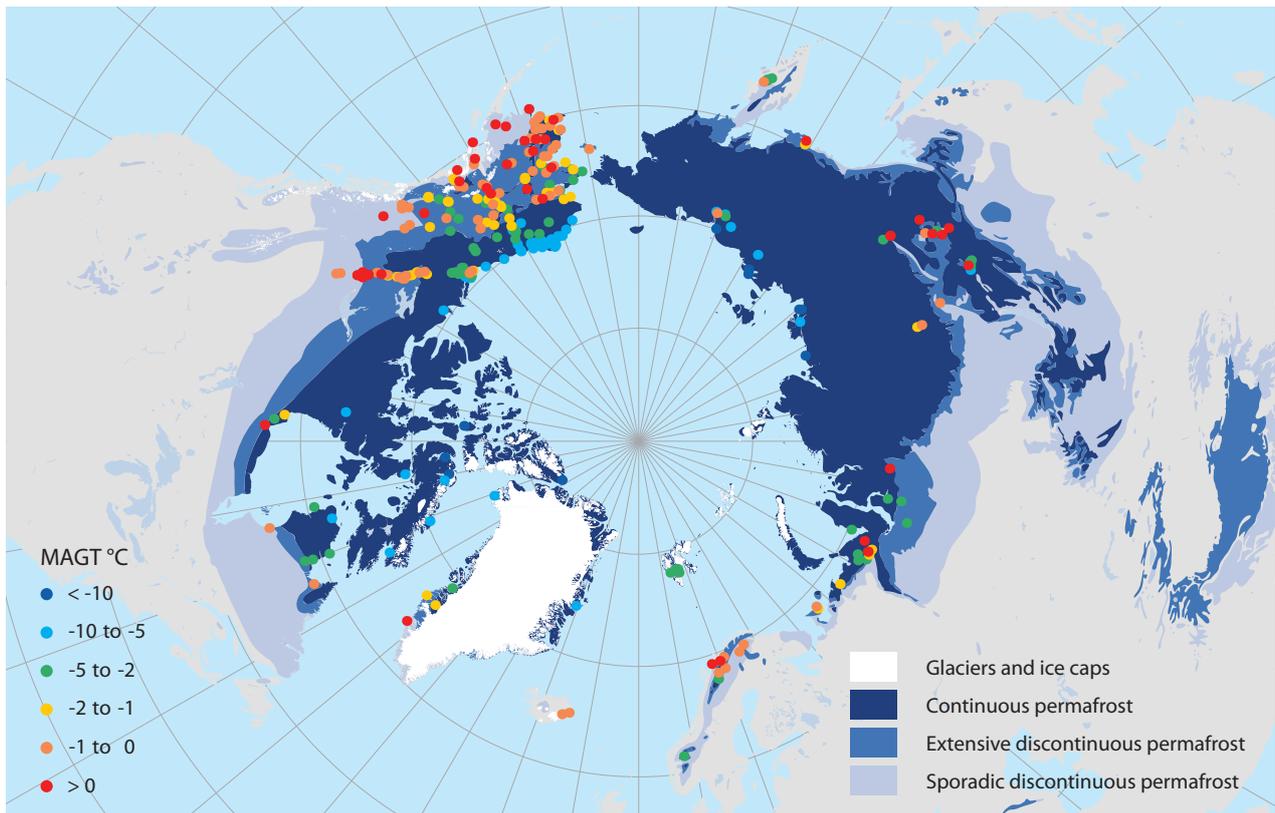
Figure 5.6. Mean annual ground temperature (MAGT) determined at the depth of zero annual amplitude, or the nearest adjacent measurement point during the International Polar Year. Source: Romanovsky et al. (2010b).

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
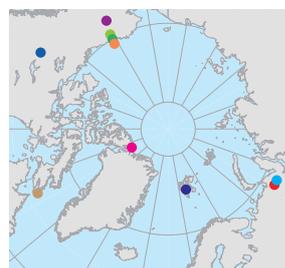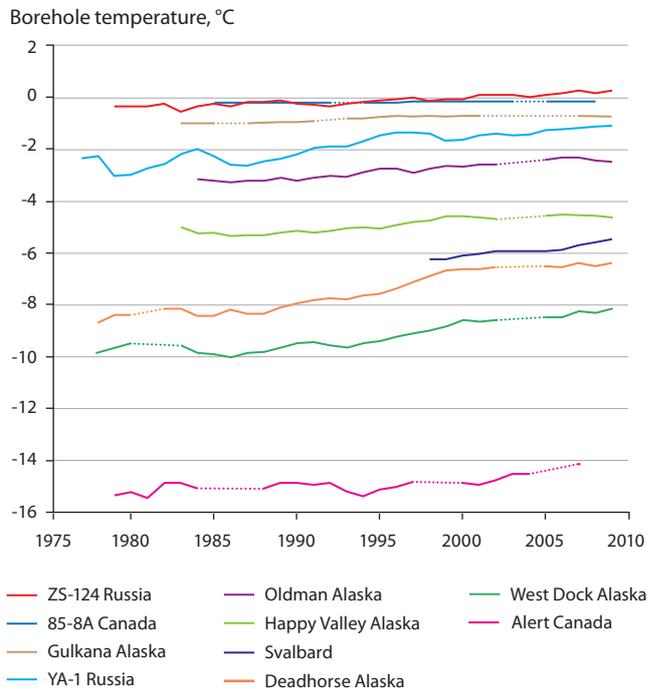
Figure 5.7. Ground temperature at depths between 10 and 20 m for boreholes across the circumpolar northern permafrost regions (Romanovsky et al., 2010b). Data sources are as follows: North America (Smith et al., 2010), Russia (Romanovsky et al., 2010a) and Svalbard (Isaksen et al., 2007b; Christiansen et al., 2010).

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
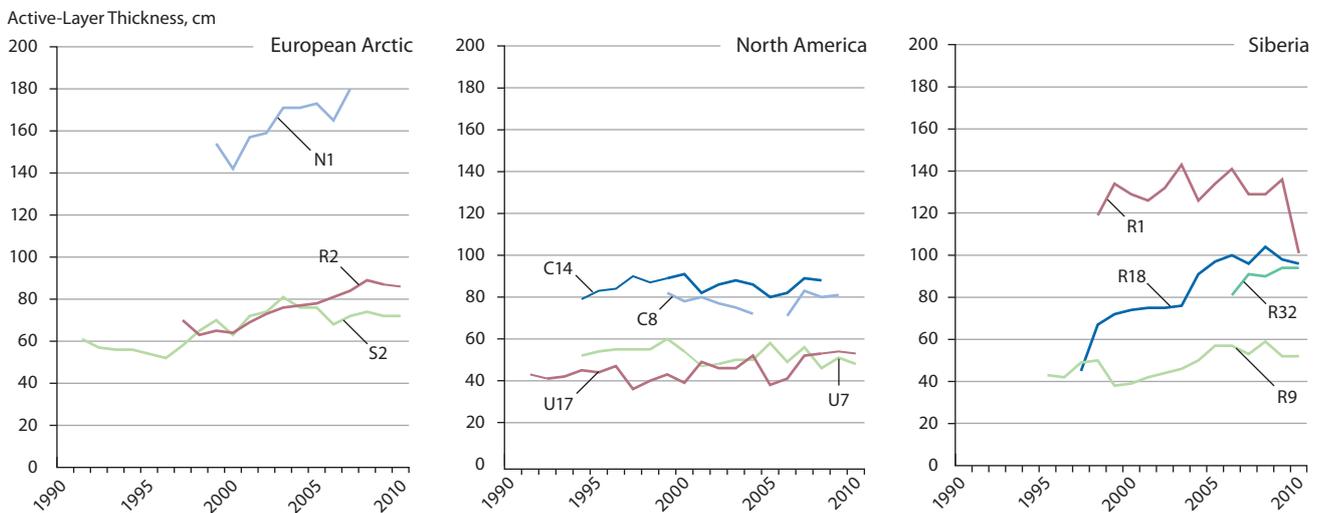
Figure 5.8. Distribution of northern hemisphere CALM sites (see Figure 5.4 for detailed permafrost map) and examples of the trends in active-layer thickness at a selection of sites (indicated by red dots). The active-layer measurements were obtained from bedrock, sediment and organic-rich sites. Site numbers correspond to the CALM nomenclature and the entire CALM data archive is available at www.udel.edu/Geography/calm/data/data-links.html.

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 result in greater increases 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
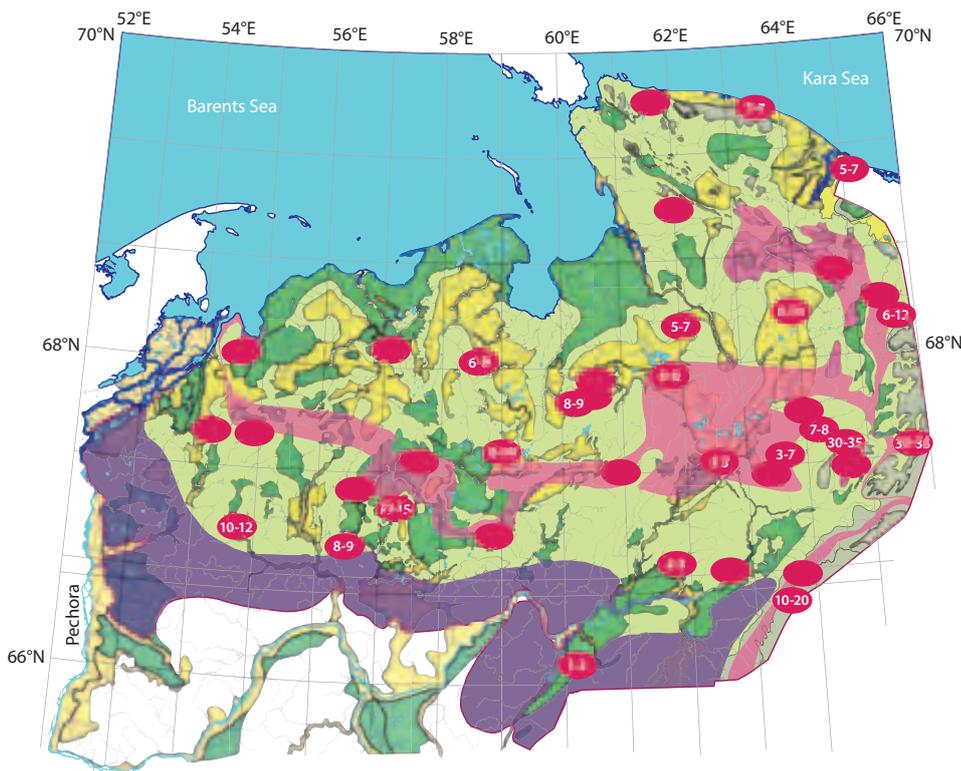
Figure 5.9. Regional changes observed in permafrost dynamics in the Russian European North over the period 1970 to 2005. Source: Oberman and and Liygin (2009).

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 previous coastal terrestrial areas are inundated, ground temperature and salinity at the current 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.

Over the past decade, several studies have projected the impacts of climate warming on permafrost in the 21st century. Most ACIA (2005) projections were 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

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 as 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.

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,
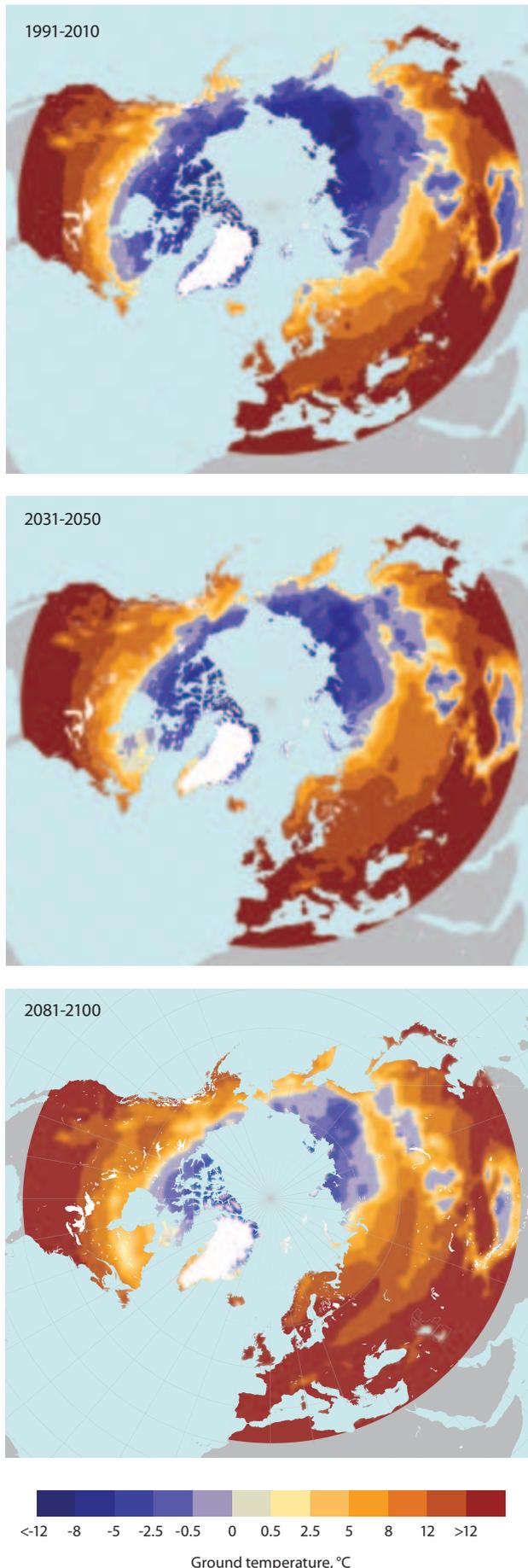
Figure 5.10. Arctic-wide projections of ground temperature for 2010 (upper), 2050 (middle) and 2090 (lower) using the GIPL-model.

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 (Figure 11 top row). 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 5.11 top right). 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 5.11 middle right). 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 (Figure 5.11 bottom row). 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
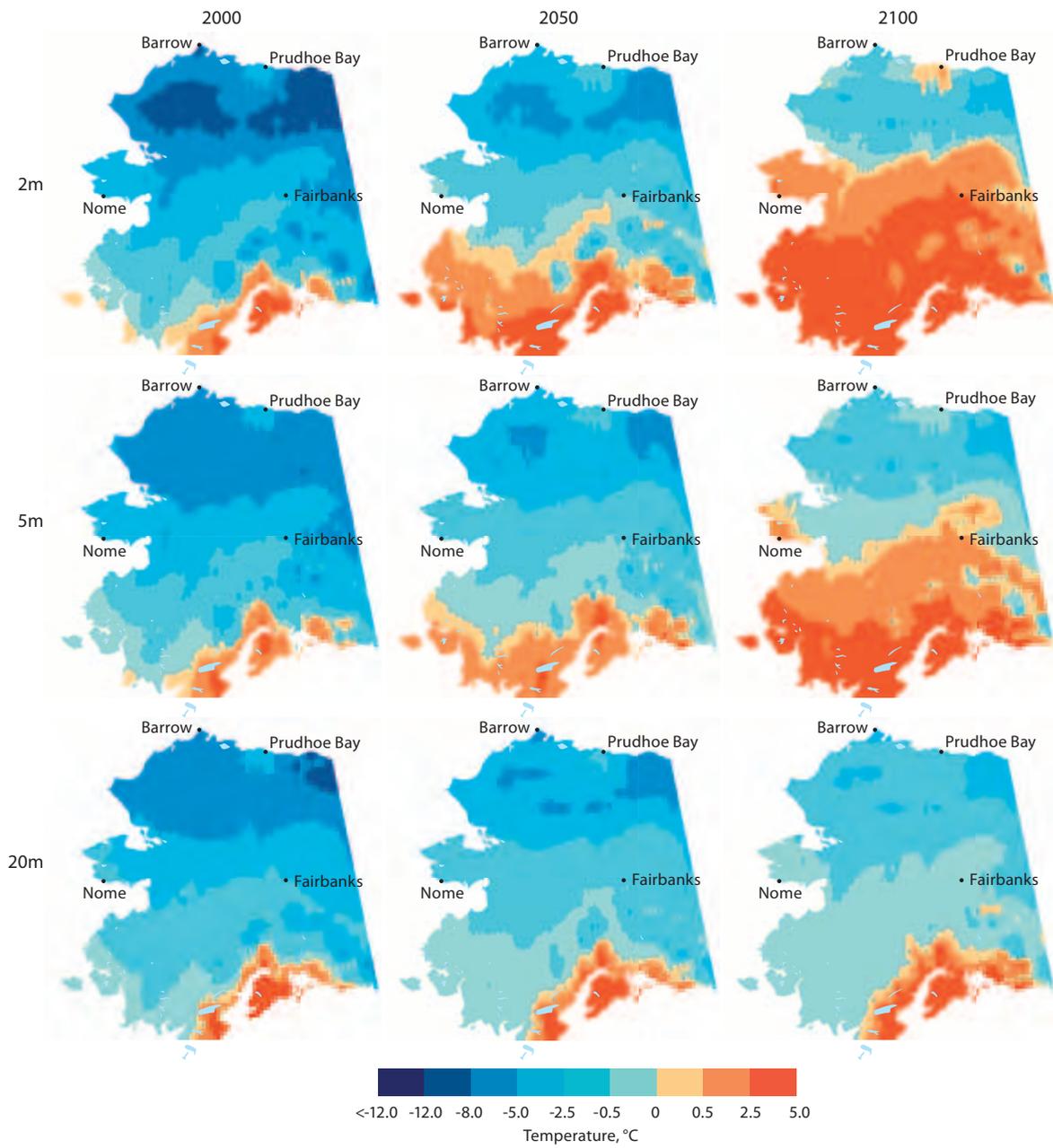
Figure 5.11. Projected mean annual ground temperature at 2 m, 5 m, and 20 m depths in 2000, 2050, and 2100 using climate forcing from MIT-2D model output for the 21st century. Source: Marchenko et al. (2008).
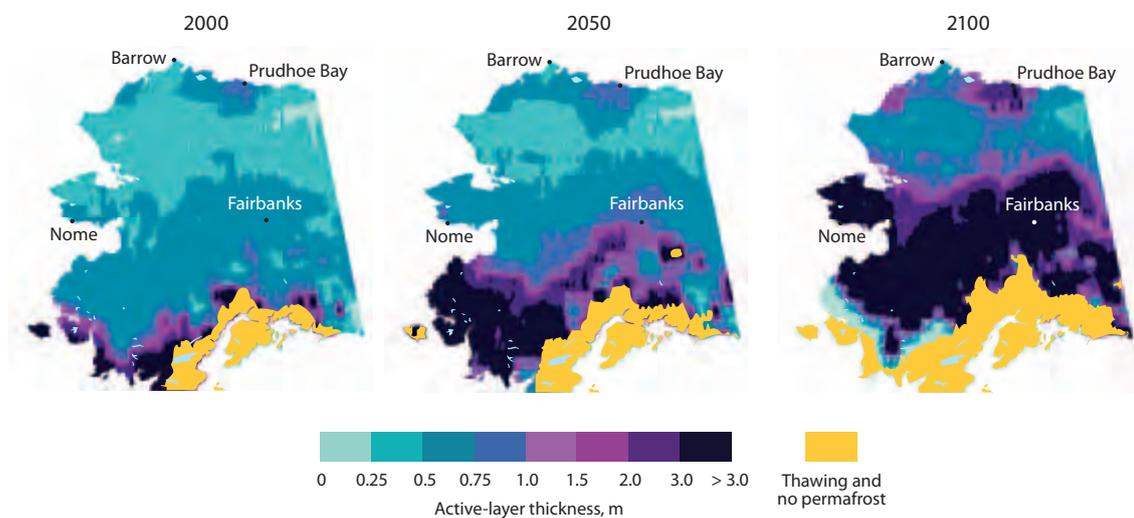
Figure 5.12. Projected active-layer thickness and extent of thawing permafrost area in 2000, 2050, and 2100 using climate forcing from MIT-2D model output for the 21st century. Source: Marchenko et al. (2008).

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 to 1990) 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 the Yukon Territory 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.
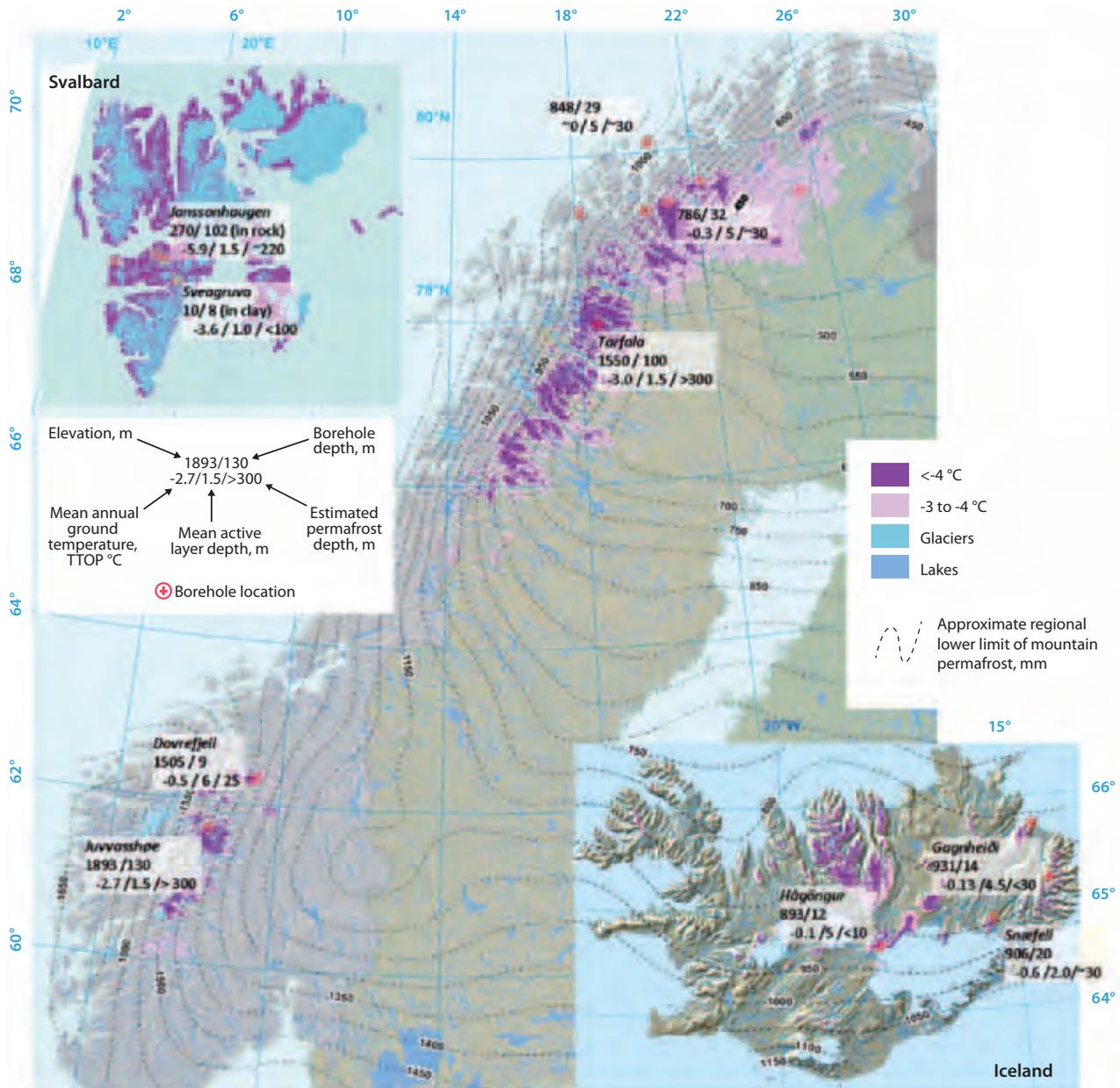
Figure 5.13. Projected changes in permafrost within the Nordic countries based on changes in air temperature. The effects of snow cover and topography are not considered. Source: compiled by the University of Oslo (UiO), Norway, based on data provided by UiO, the Norwegian Meteorological Institute, the University Centre in Svalbard, and the Geological Survey of Norway.

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 of 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.
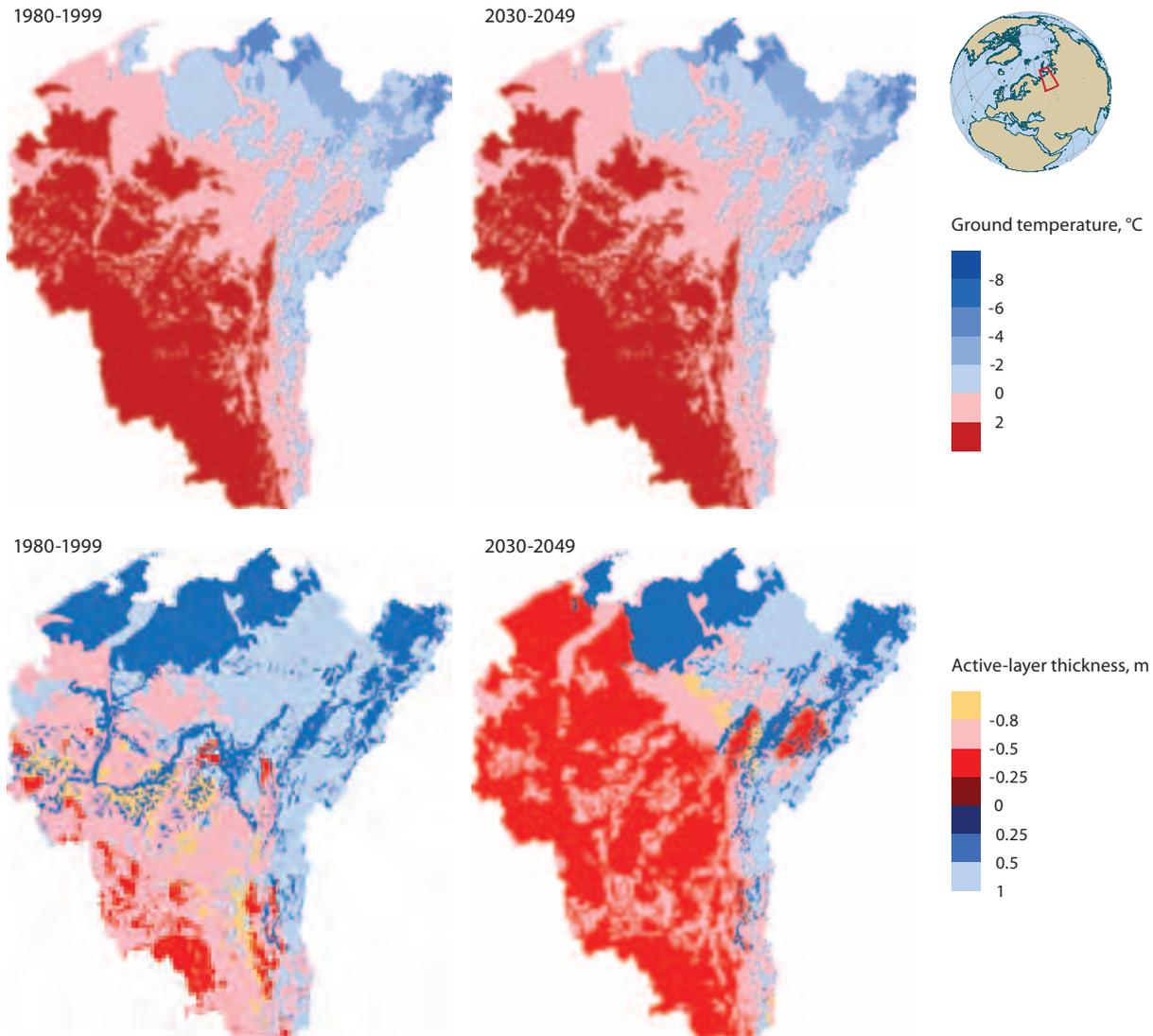
Figure 5.14. Modeled mean annual ground temperature for the Pechora River catchment, Russia, at the bottom of the active layer (upper) and active-layer thickness and seasonal frost penetration (lower) for 1980 to 1999 (left) and averaged for 2030 to 2049 (right). The results are derived from the GIPL-1.3 equilibrium permafrost model run using the 4 × 4 km HIRHAM5 climate scenario. Source: Stendel et al. (2010).

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
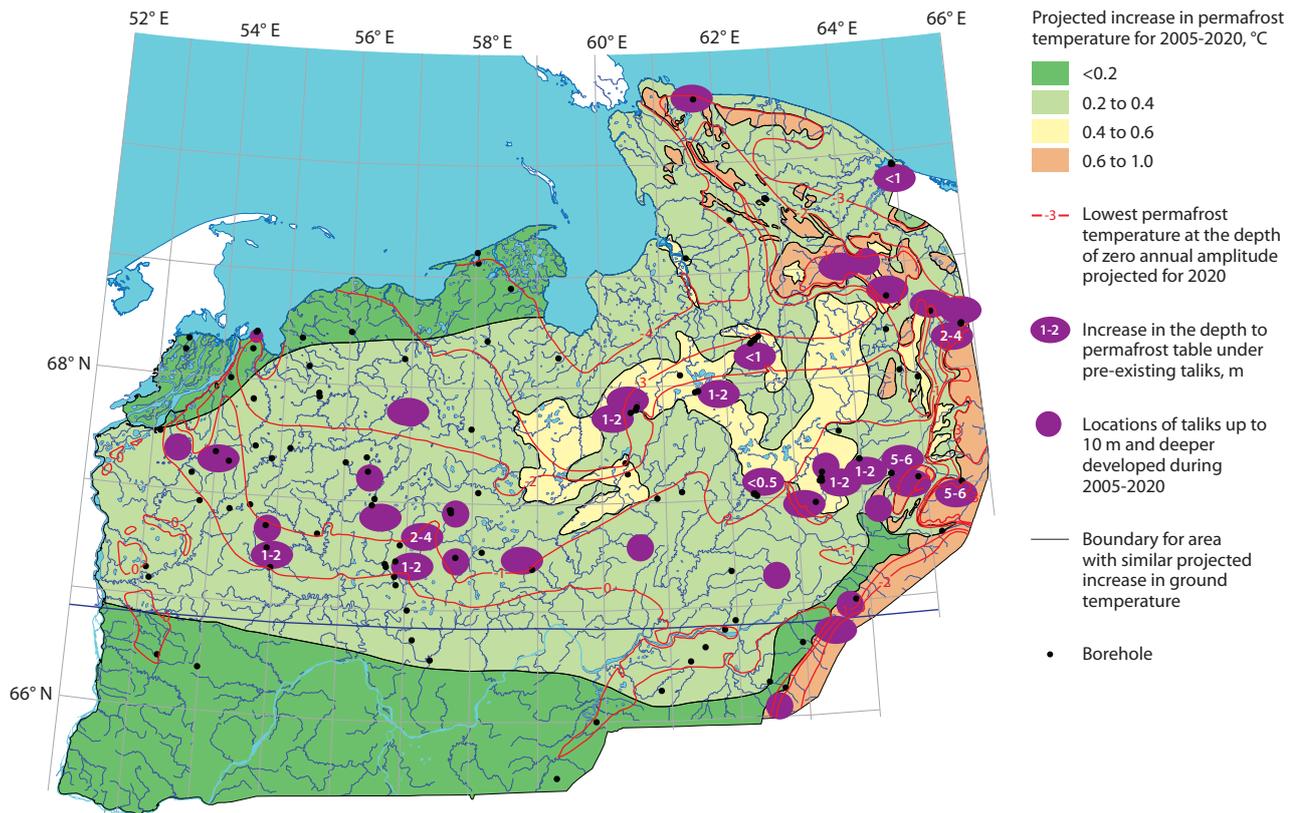
Figure 5.15. Changes projected in permafrost by 2020 in the Russian European North. Source: Oberman and Liygin (2009).

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 (Mack et al., 2011). 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; Bunn 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
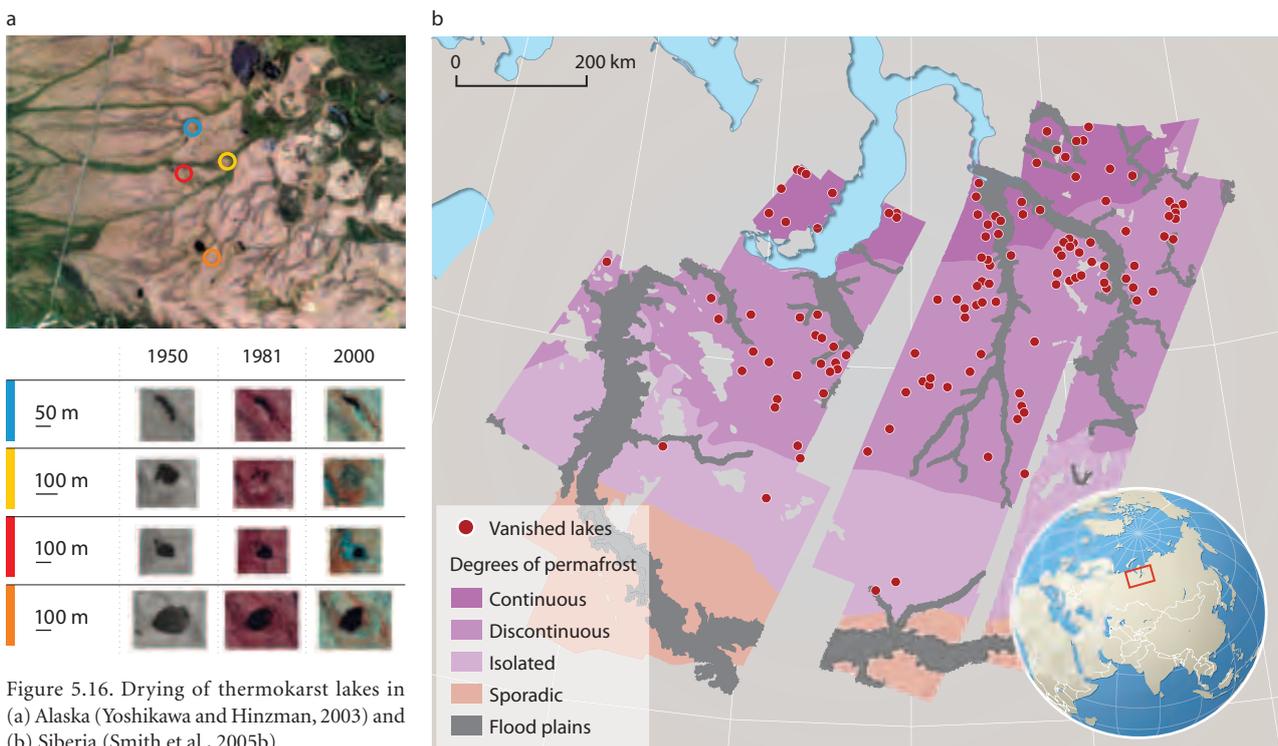
Figure 5.16. Drying of thermokarst lakes in (a) Alaska (Yoshikawa and Hinzman, 2003) and (b) Siberia (Smith et al., 2005b).
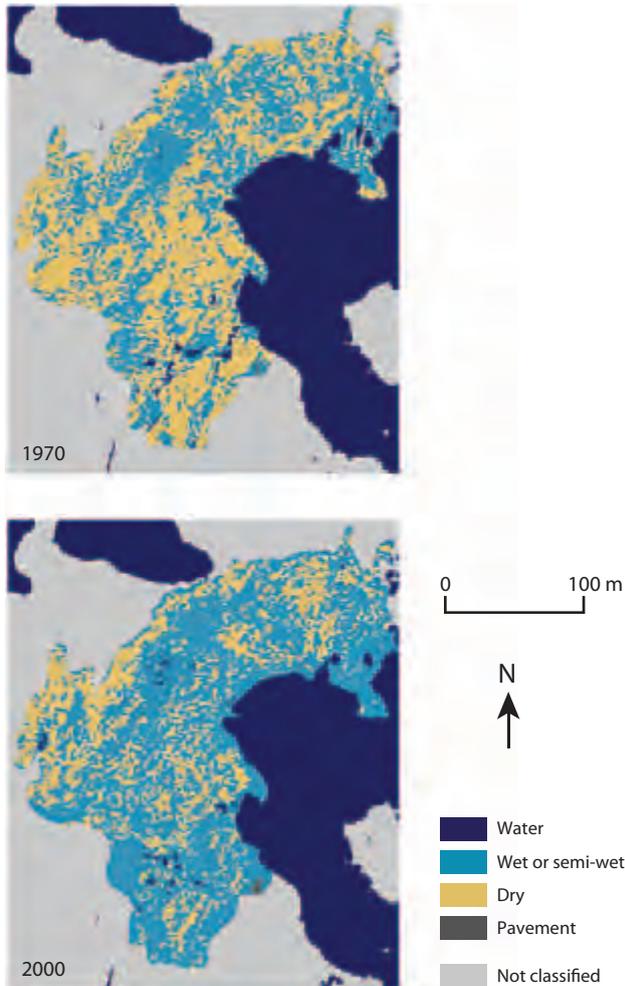
Figure 5.17. Waterlogging in areas with thawing permafrost in sub-Arctic Sweden – a comparison between the situation in 1970 (upper plot) and in 2000 (lower plot). Source: Christensen et al. (2004).

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; Gooseff 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).
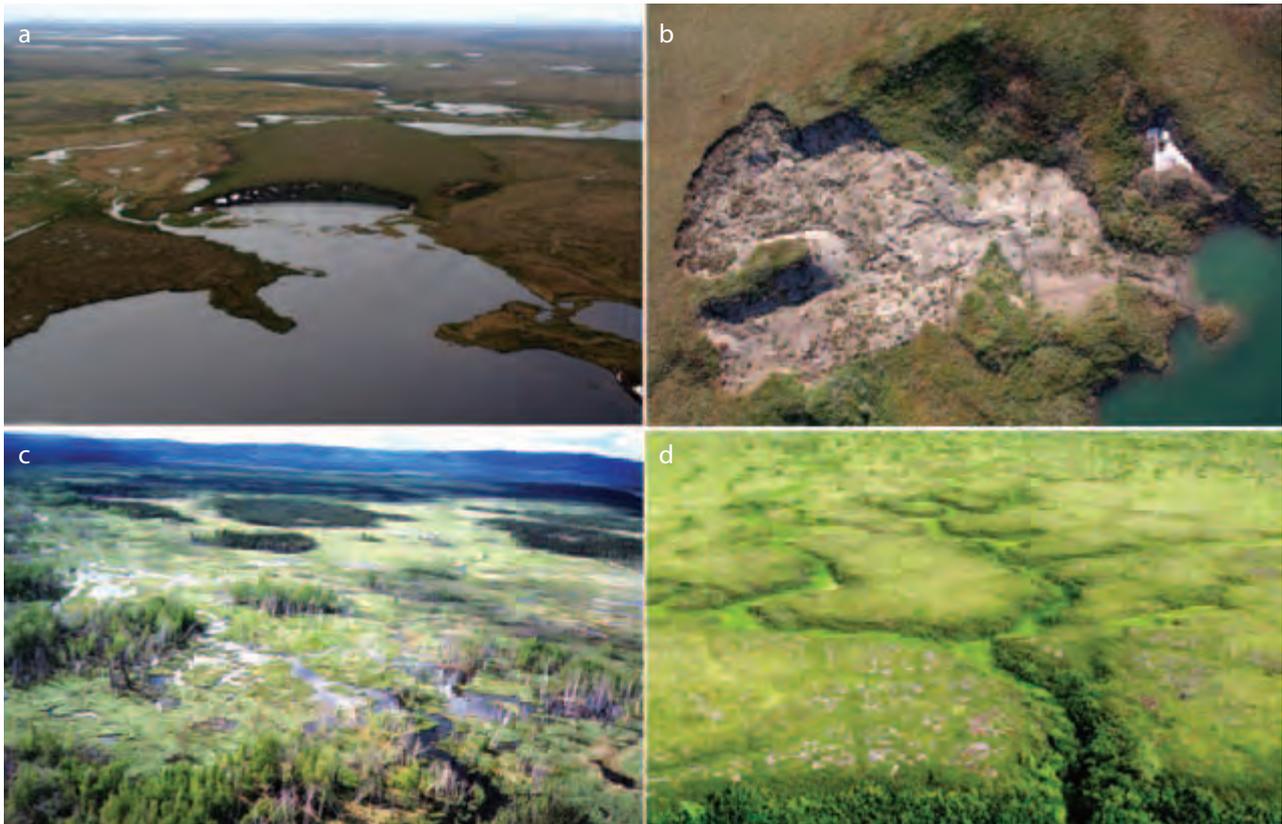
Figure 5.18. Common thermokarst landforms, including (a) thermokarst lakes, (b) retrogressive thaw slump associated with glacial thermokarst, (c) thermokarst fens and bogs, and (d) thermo-erosional gullies. Photographs by M.T. Jorgenson.

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

Figure 5.20. Recently eroded tunnel and disturbed topography developed by meltwater from thawing of ice-wedges on Bylot Island, Canada. Photo Michel Allard.

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).

Figure 5.19. Ice-wedge polygons. Photo: Ulrich Neuman, Kolibri Geo Services

Figure 5.21. Four open system pingos in different stages of development at Reindalen, Svalbard, April 2008. The tallest pingo is 30 m high, the closest is 100 m in diameter. Photo: Hanne H. Christiansen.

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
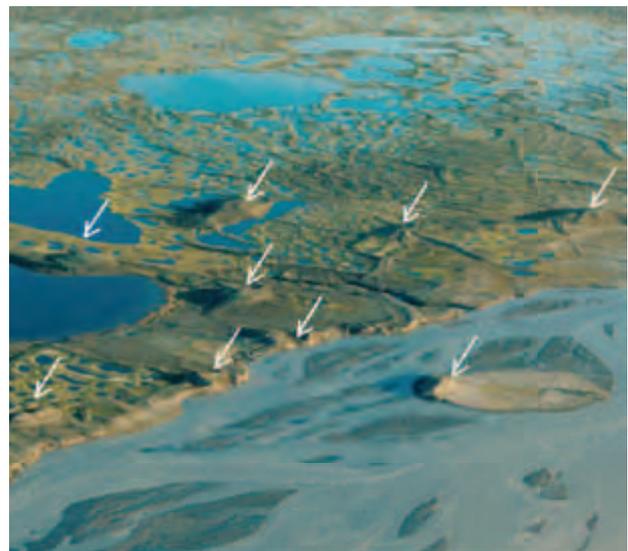
Figure 5.22. Pingos (indicated by arrows) in ice-wedge polygon terrain on a terrace covered with thaw lakes on Bylot Island, Canada. The eroded pingos along the riverbank and on the island are collapsing. Photo: Michel Allard.

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 those 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,
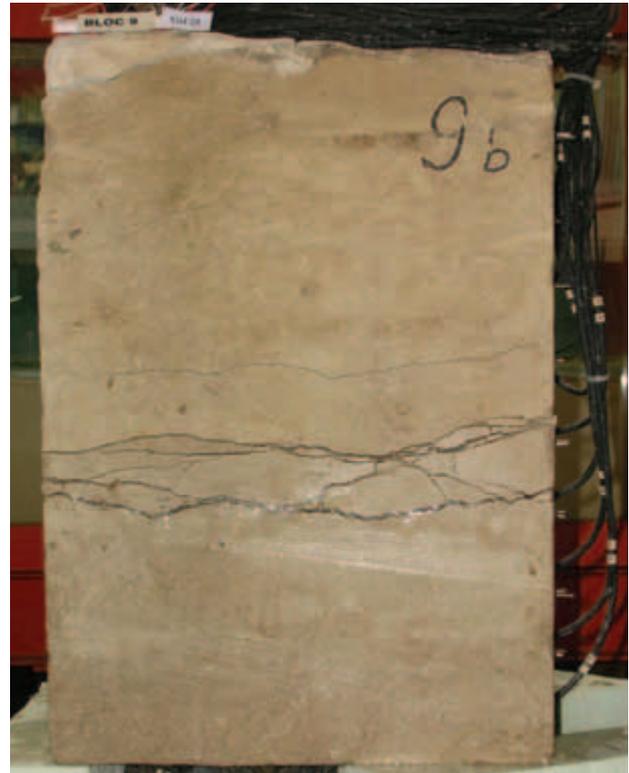
Figure 5.23. Fractures formed by ice segregation in Totternhoe Stone (a type of English chalk) following twenty-four cycles of bidirectional freezing of an active layer above permafrost. The fractures are filled with segregated ice and occur primarily in near-surface permafrost and at the base of the active layer. Some narrow cracks are also visible in the upper part of the active layer, caused by downward freezing from the rock surface. The block is 450 mm high and 300 mm wide. Photo Julian Murton, University of Sussex, UK.

Figure 5.24. Rock glaciers creeping from a cirque (left) and directly from a free face (right) on Disko Island, central West Greenland. The rock glacier to the left is about 1100 m long. The mountains are about 900 m high. Photo: Ole Humlum, University of Oslo, Norway, 8 July 1999.

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 m³) 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

Figure 5.25. Talus accumulation produced by rockfalls and debris flows. Endalen, central Svalbard. Photo: Ole Humlum, 5 August 2000.

Figure 5.26. Opening of a 10 to 15 m deep crack in the rockslide area, Nordnes, northern Norway, with the Storfjorden below, September 2005. Photo: Hanne Christiansen, UNIS, Svalbard.

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 becomes thicker. 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 thicker 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
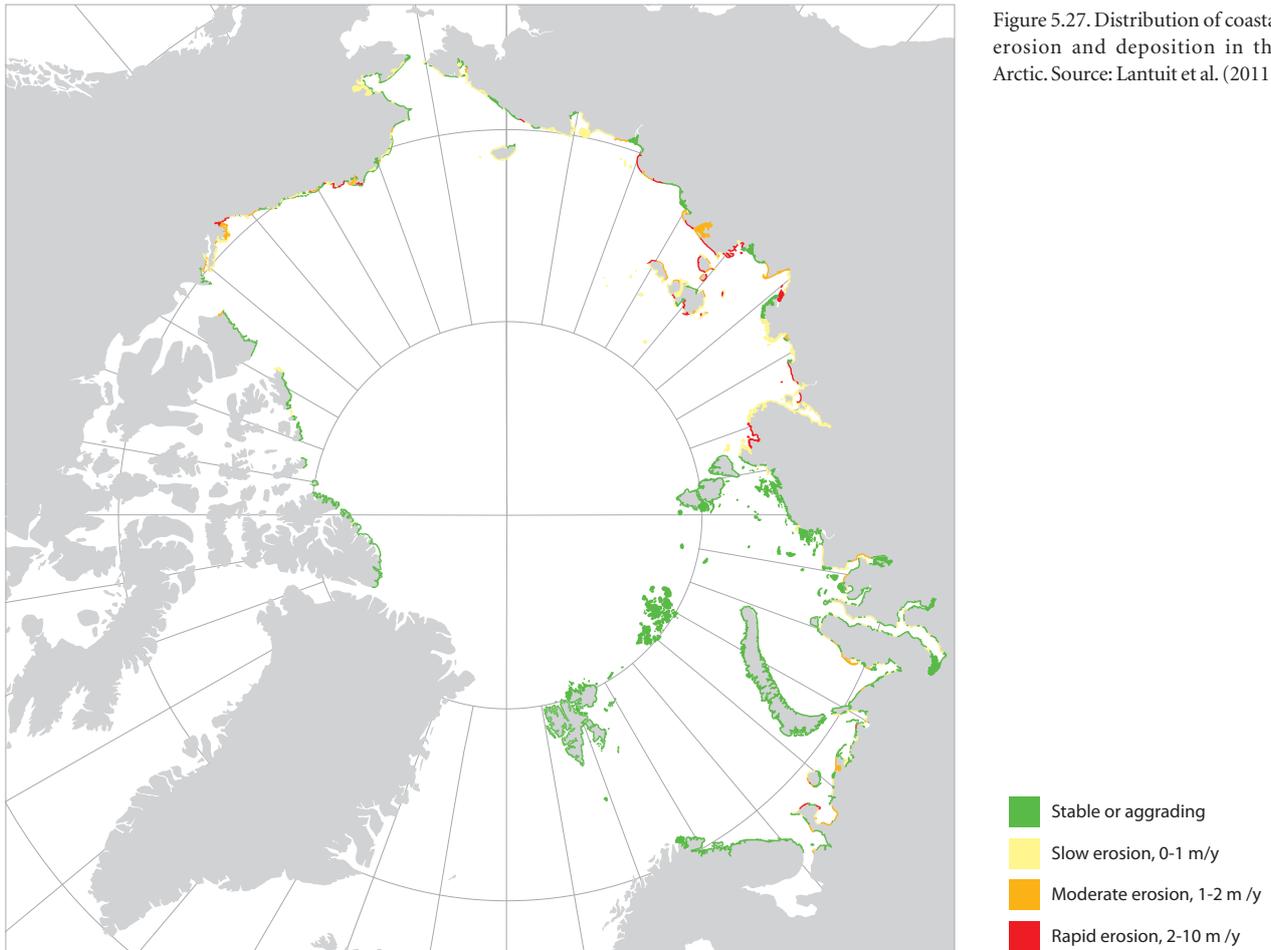
Figure 5.27. Distribution of coastal erosion and deposition in the Arctic. Source: Lantuit et al. (2011).

Stable or aggrading

Slow erosion, 0-1 m/y

Moderate erosion, 1-2 m /y

Rapid erosion, 2-10 m /y

al., 2007; see also Chapter 11, Section 11.2). 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 satellites (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).

Figure 5.28. 'Drunken forest' occurs when the ground becomes unstable due to thawing permafrost. Photo: Trofim Maximov, Institute for Biological Problems of Cryolithozone, Russian Academy of Sciences, Yakutsk, Russia.

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 km², which was more than the sum of all known fires on the North Slope since 1950 (Jones et al., 2009; Mack et al., 2011; 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 10^7 cells 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

Figure 5.29. Tundra fire on the North Slope of Alaska in 2007 resulted in thawing of the upper 5 to 10 cm layer of permafrost. Photo: Michelle Mack.

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 relict 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 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/m² per decade. Complete conversion of tundra to tree cover in northern Alaska is estimated to increase summer heating by around 26 W/m² (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 4.3.1.3.2). 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 Chapter 11, Section 11.1 on physical feedbacks and forcings). 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), 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; Grøndahl 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

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 /m², 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.

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 5.30). 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.
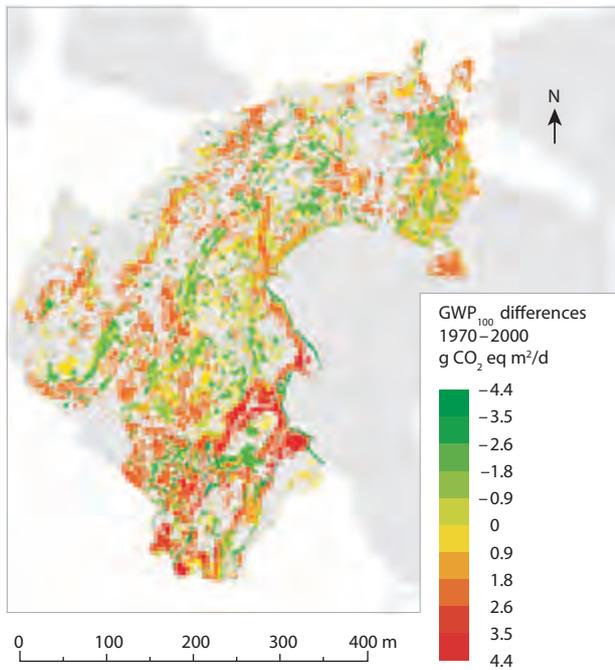
Figure 5.30. Spatial distribution of changes in annual net exchange of carbon dioxide (CO_2) equivalents for the period 1970 to 2000, where both CO_2 and methane (CH_4) are incorporated, Stordalen, northern Sweden. GWP_{100} represents Global Warming Potentials over 100 years. Source: Johansson et al. (2006).

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).

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).

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

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).

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.

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 permafrost 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.

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, 2011; Mack et al., 2011). 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., 2010; 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 5.31). 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 (≤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
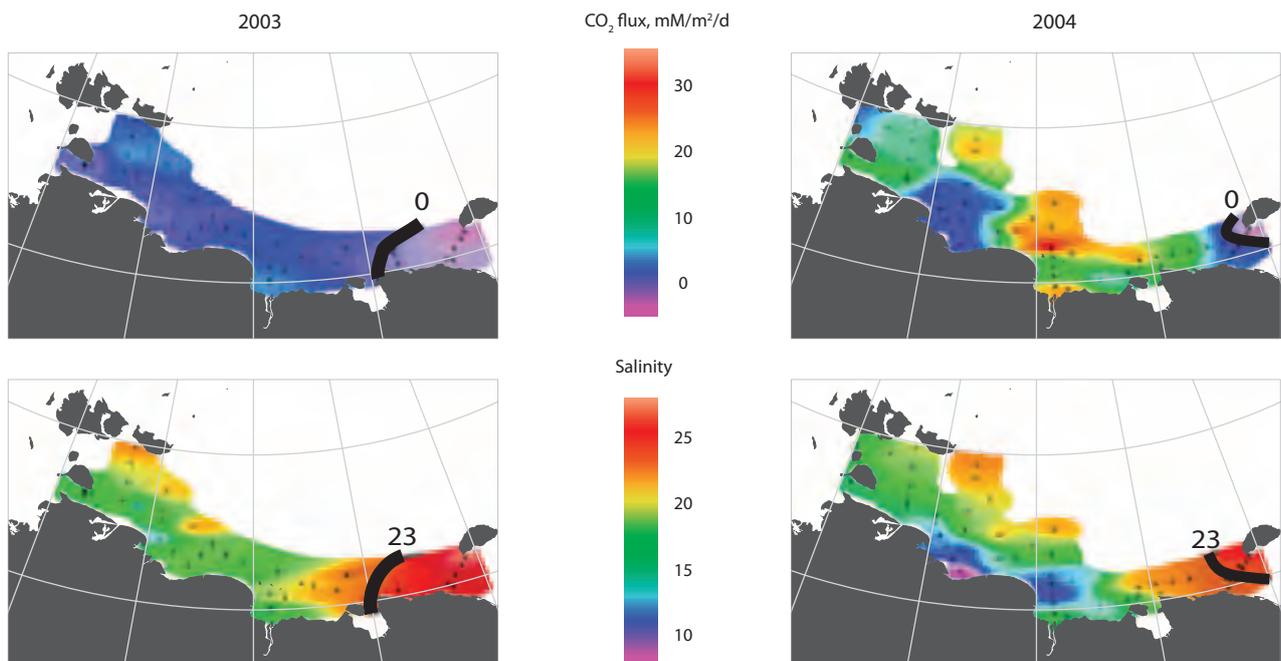
Figure 5.31. Air-sea CO_2 fluxes and surface salinity in the East Siberian Sea in 2003 (left), and in 2004 (right – to show interannual variability) (modified from Semiletov et al., 2007). Bold black lines (0-line) mark change in flux direction, which agrees with the position of the hydrological frontal zone (salinity 23 line) (modified from Semiletov et al., 2005). The predominant substrate for CO_2 production comes from coastal erosion of terrestrial organic carbon in retreating ice-complexes.
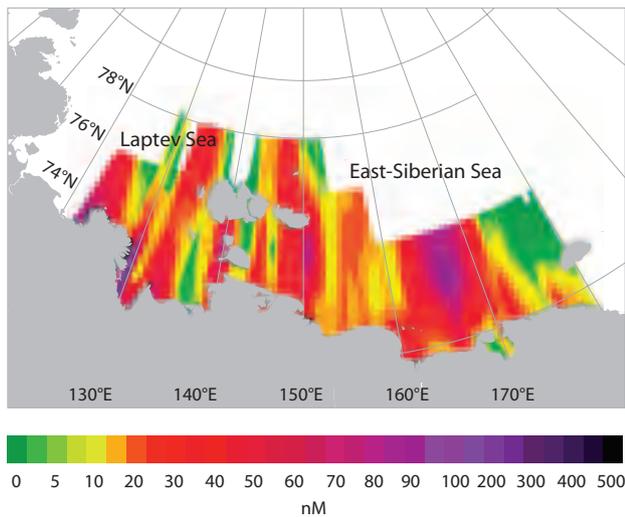
Figure 5.32. Dissolved methane distribution in the surface water of the East Siberian Arctic Shelf for the period 2003 to 2008. Source: Shakhova et al. (2008a).

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 5.33a), 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 5.33b). 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 km²) 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
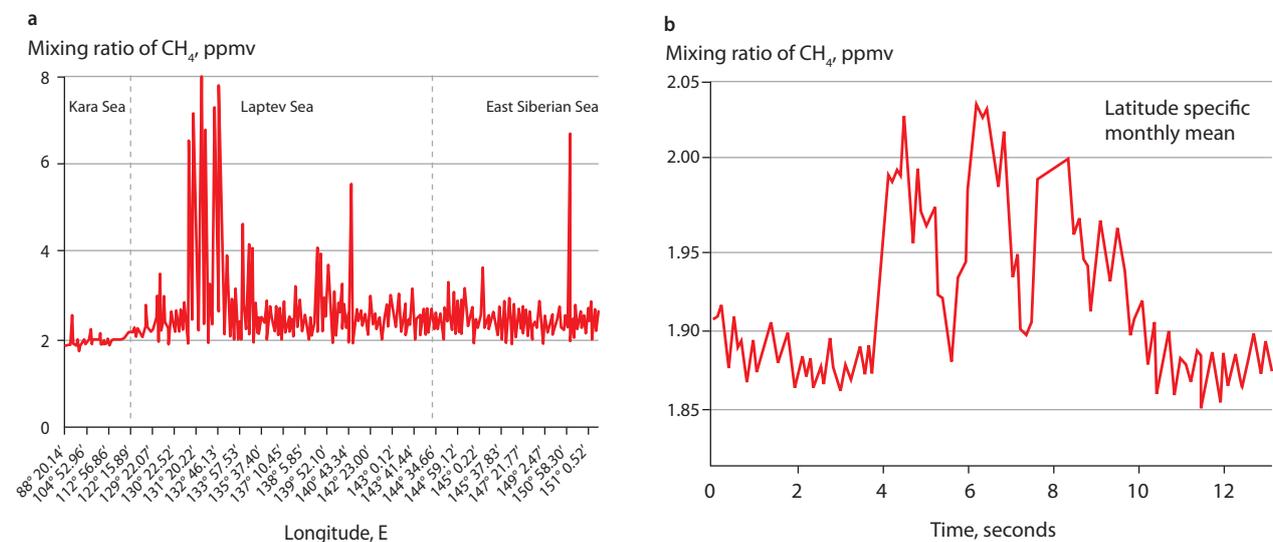
Figure 5.33. Methane release from the East Siberian Arctic Shelf: (a) mixing ratio of methane in the atmospheric layer above the sea surface measured along the ship route in September 2005 (b) pulse character of the methane release. Source: Shakhova et al. (2008b).

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 and 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).

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.

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

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		

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

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 5.34).

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.
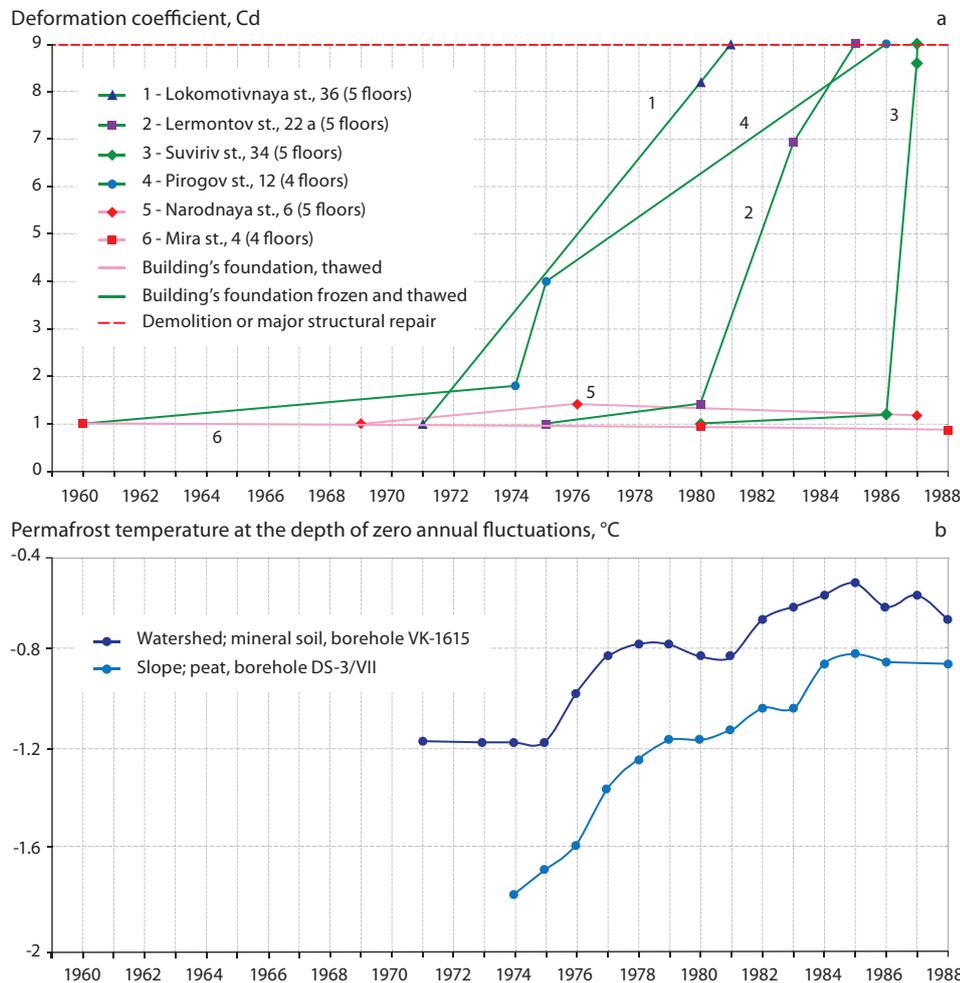
Figure 5.34. (a) Deformation coefficients used to evaluate structural integrity of multi-storey buildings in Vorkuta in northwestern Russia (Oberman, 2003a). A deformation coefficient (C_d) of 1, means that no deterioration will take place. Increasing C_d indicates increasing deterioration of the building and a value of 9 means that major repairs or demolition of the building is necessary. (b) Catastrophic deformations of buildings are remarkably confined to only one of three decades (i.e. to the 1980s). These are the years characterized by the greatest increase in temperatures of permafrost composed of mineral Quaternary deposits and peat.

Mean annual air temperature, °C

Maximum winter snow depth, cm
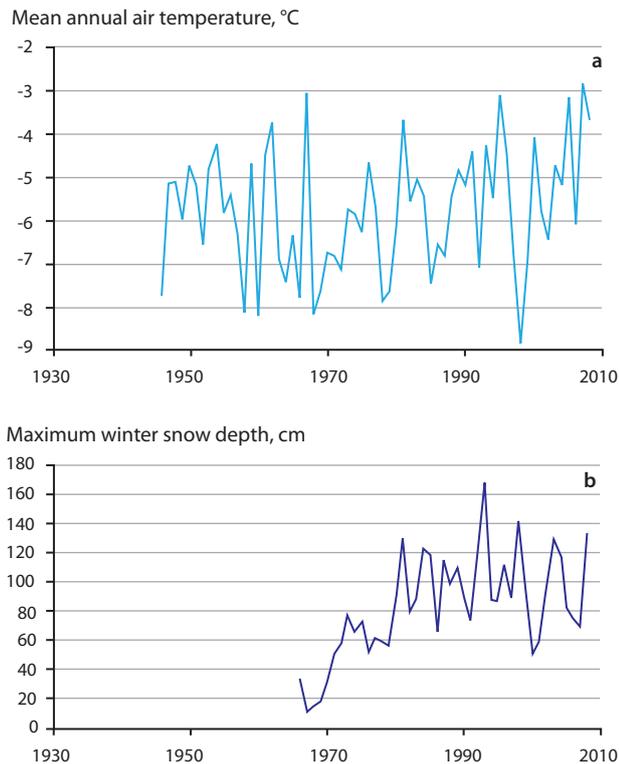
Figure 5.35. Mean annual air temperature and maximum winter snow depth recorded at the Vorkuta meteorological station, northwestern Russia. Source: Romanovsky et al. (2010a).

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., 2008a), 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

Borehole temperature, °C
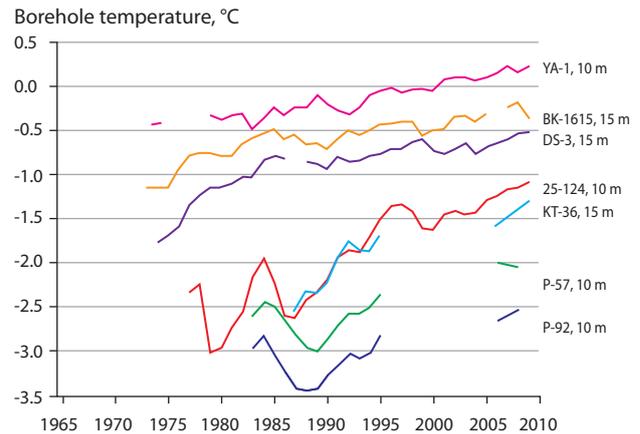
Figure 5.36. Time series of ground temperatures at 10 and 15 m depth at different locations within the Vorkuta research region. All the boreholes numbered above have the following landscape description: ZS-124 – peatland with ice wedge polygons; KT-36 flat surface of an alluvial-marine terrace; P-92 – gently sloping surface of an alluvial-lacustrine plain; P-57 – flat surface of an alluvial-lacustrine plain; BK-1615 – catchment of a glacial-marine plain; DS-3 – ice wedge polygons on a glacial-marine plain; YA-1 – within a water track on an erosional plain. Source: Romanovsky et al. (2010a).

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 amounts of fossil fuels and is an important supplier of oil and gas to the global market (Table 5.3).

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

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

change. However, climate change effects on permafrost will possibly affect those maintenance, 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. Examples are 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 projections 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'
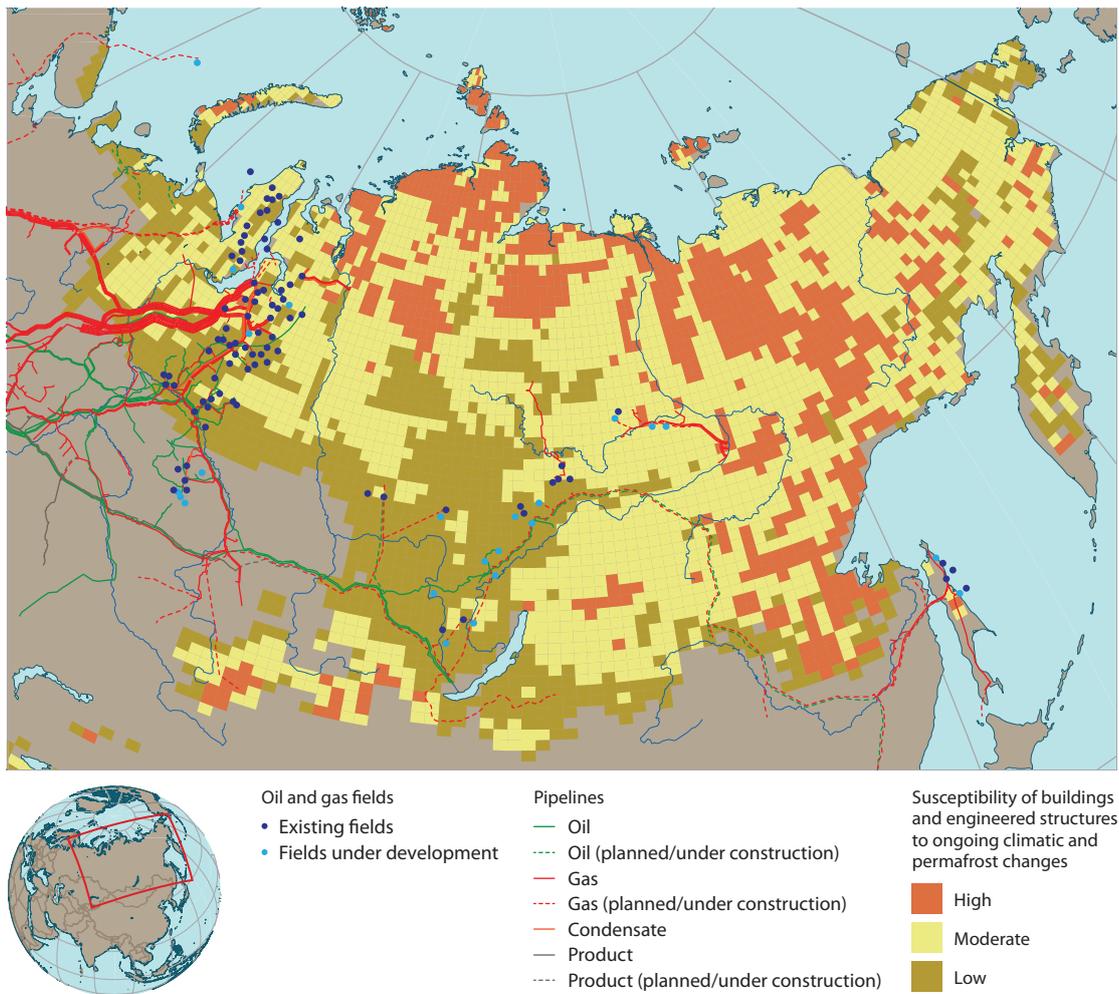
Figure 5.37. Predictive map of potential threats to infrastructure for 2050 in Russian permafrost regions. Colors denote regions with high, moderate, and low susceptibility of buildings and engineered structures to ongoing climatic and permafrost changes (updated from Instanes and Anisimov, 2008). Compared to the original publication, areas with isolated permafrost islands within the sporadic zone have been eliminated using the IPA digital map of permafrost and ground ice conditions. Major oil and gas pipelines and oil and gas fields have been added.

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).

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.

Probability of active-layer thickness exceeding threshold limits
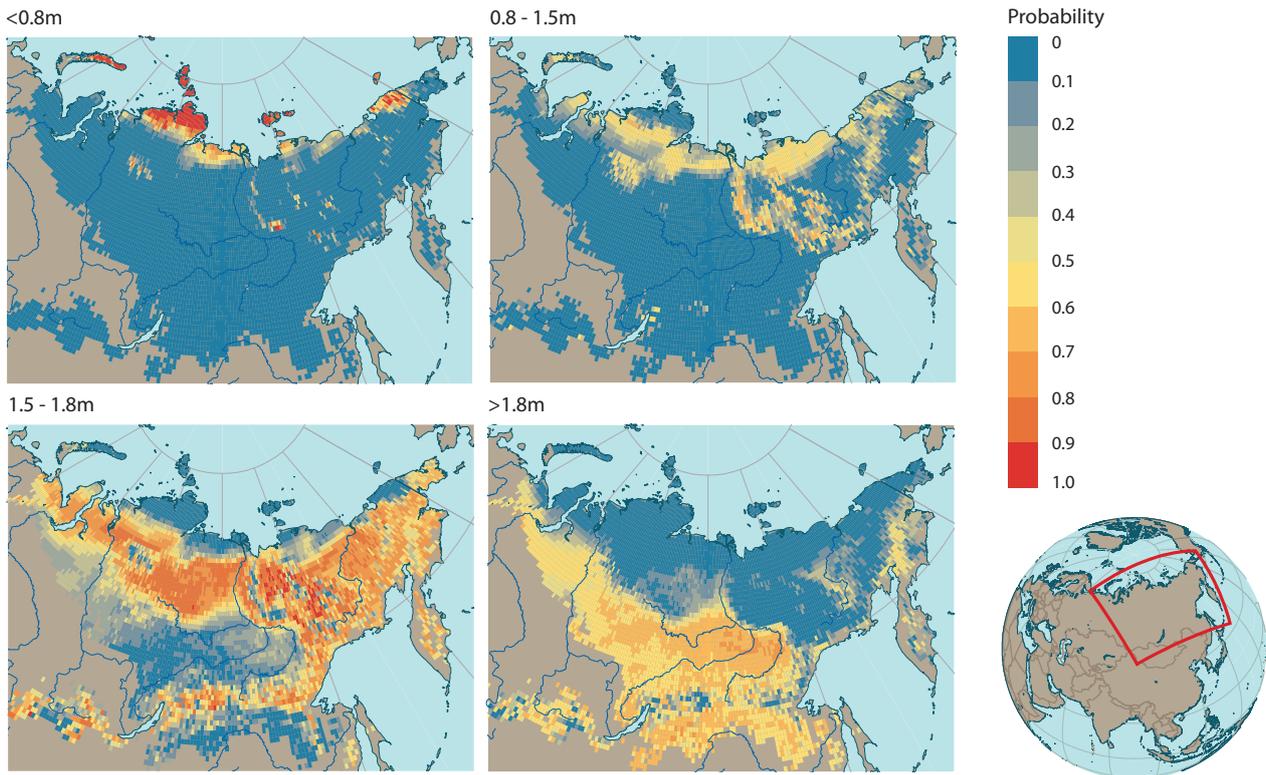
Figure 5.38. Probabilistic map of active-layer thickness for Eurasia under current climatic conditions. The scale shows probabilities of active-layer thickness exceeding the threshold limits assigned to each panel. Source: Anisimov (2009).

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.

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.

Table 5.4. Climate change impact and related engineering problem.

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

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

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.

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 helps ensure that impacts related to hydrocarbon 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).

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.

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. **Recent advances in understanding**

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 years 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.

Acknowledgments

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
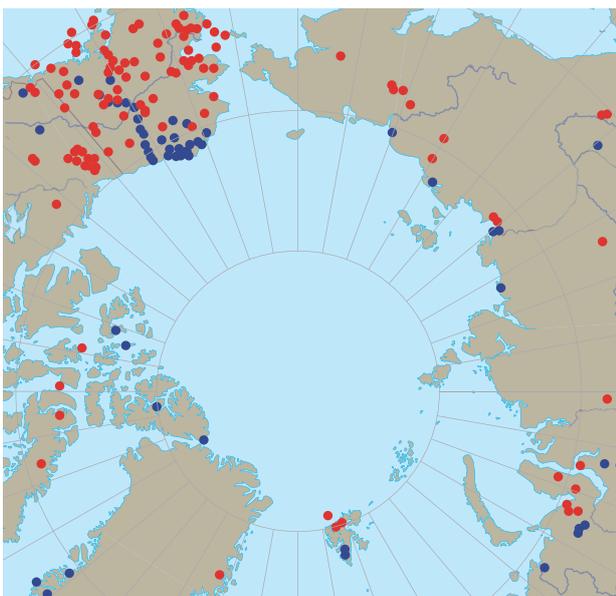
associated with the IPY are also being developed by various national organizations (see for example, Smith et al., 2008b; Juliussen et al., 2010).

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, cryosols, 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, Juliussen et al., 2010; 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 (Juliussen et al., 2010).

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,


Figure 5A.1. Long-term monitoring sites (blue dots) and sites that were established or where measurements were resumed during the International Polar Year (red dots). The sites are superimposed on the permafrost zones. Source: Romanovsky et al. (2010b).

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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Contents

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 lakes, 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
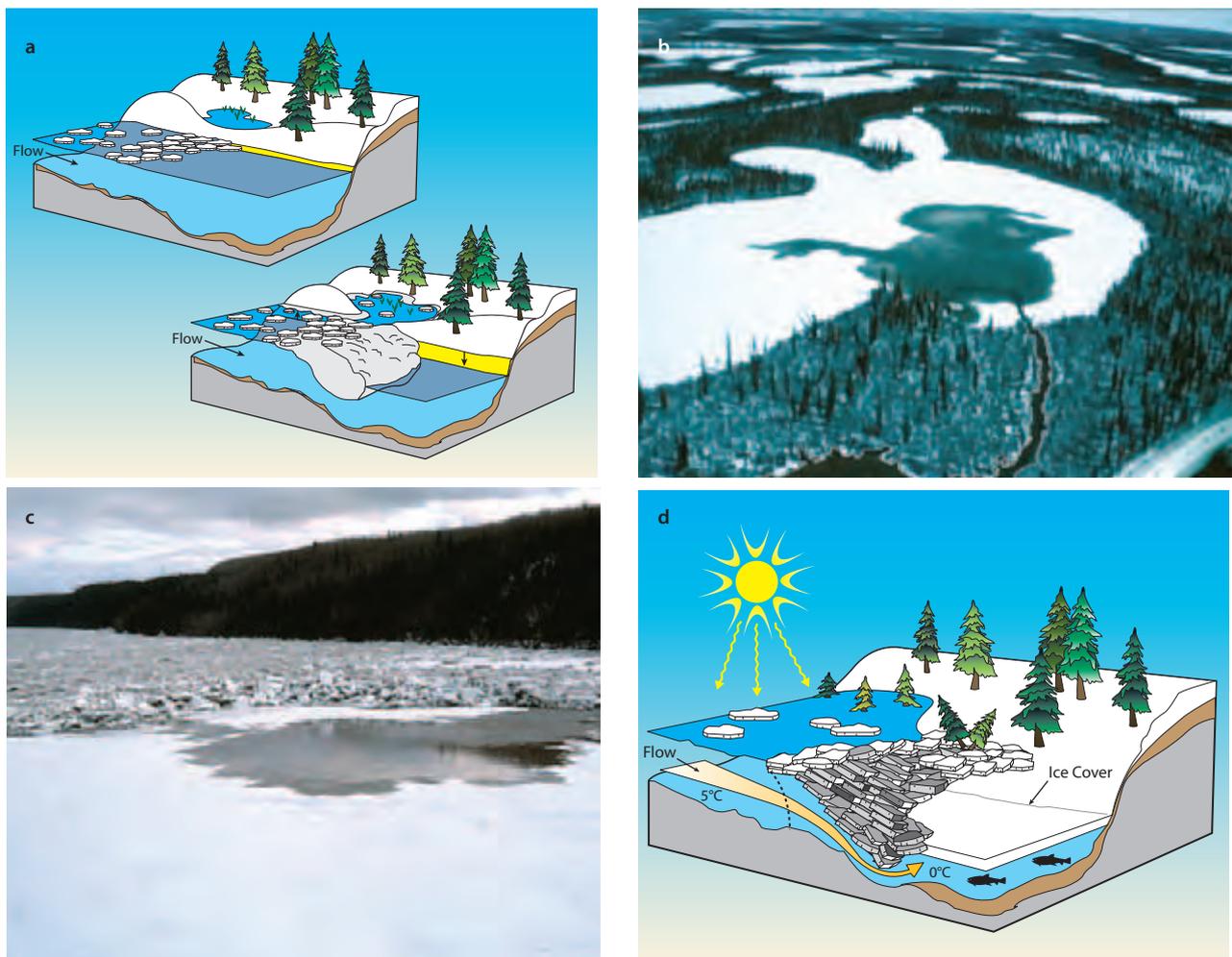
Figure 6.1. Various stages of river-ice development: (a) a schematic illustration of dynamic ice formation on rivers leading to freeze-up (water levels decrease downstream of the ice accumulation and increase upstream with 'backwater' storage); (b) backflow from the adjacent river channel during break-up feeding into a connected ice-covered lake; (c) a rapidly advancing river ice break-up front; and (d) a schematic illustration of a dynamic break-up front showing the rapid transition in water temperature, enhanced thickness of break-up accumulation, and elevated backwater, which can produce flooding. Modified from Prowse (2000).

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 km³ (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
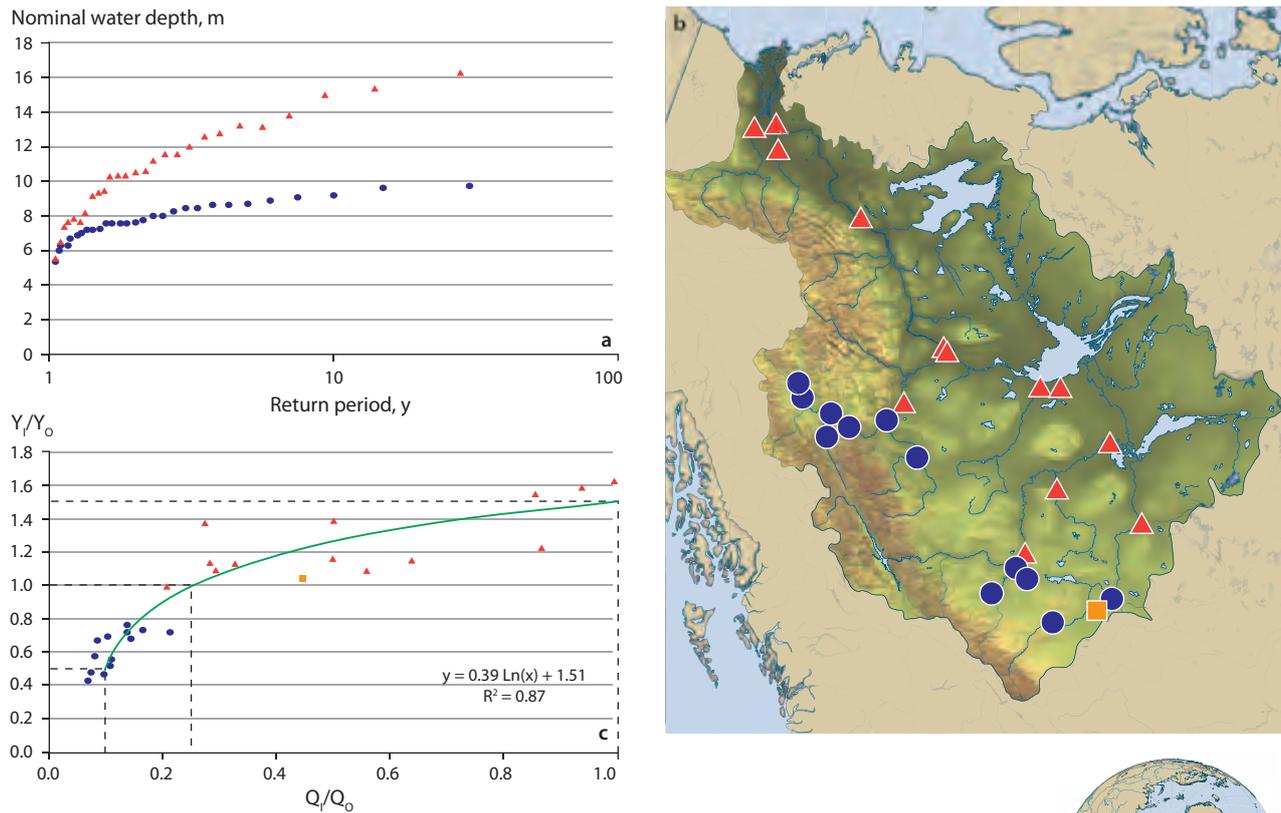
Figure 6.2. (a) Example of a return-period plot for a station on a major tributary to the Mackenzie River, Canada, in which break-up events (triangles) become increasingly important with longer return periods in producing annual high water levels as compared to open-water events (circles); (b) locations of ice (triangles), open-water (circles) and mixed (square) high-stage regimes in the Mackenzie River basin; and (c), for these regime classifications, the related dimensionless discharge (Q) versus water level (Y) curve for ice ($_I$) and open-water ($_O$) conditions. After de Rham et al. (2008a).

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 50° 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;
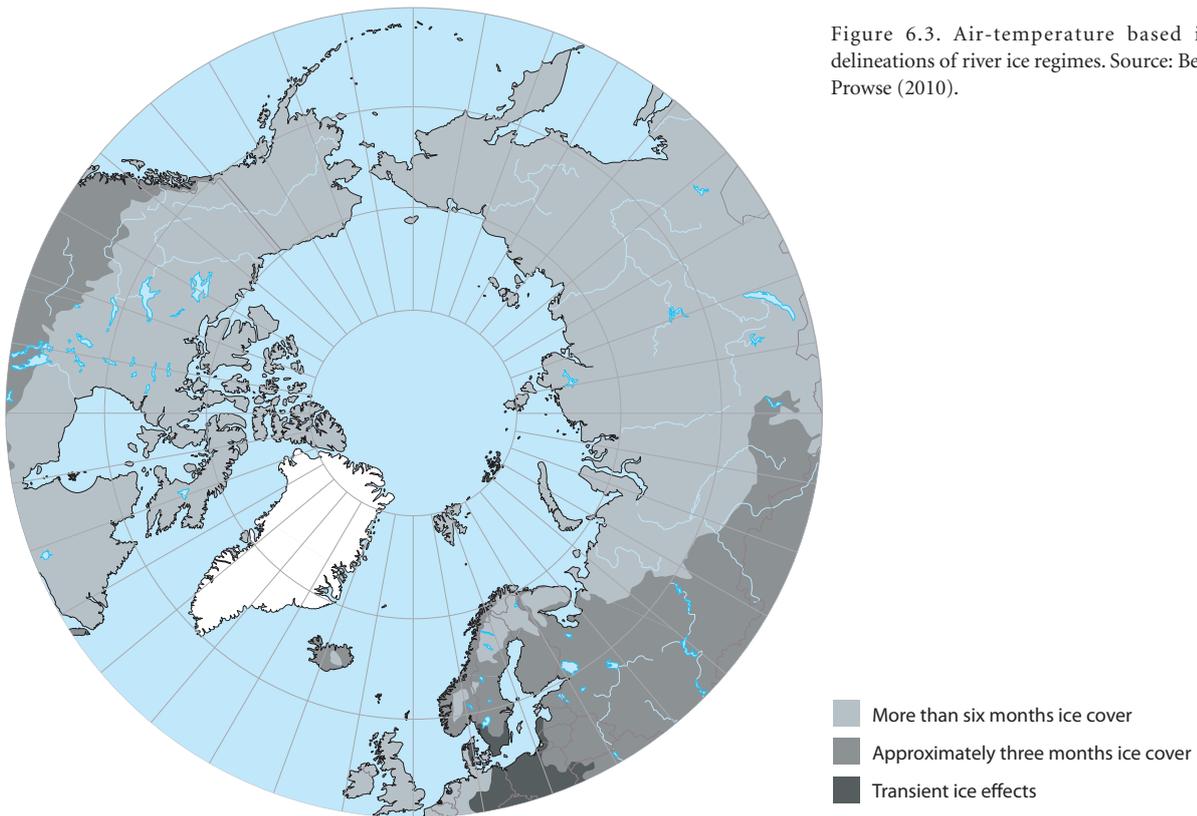
Figure 6.3. Air-temperature based isotherm delineations of river ice regimes. Source: Bennett and Prowse (2010).

More than six months ice cover

Approximately three months ice cover

Transient ice effects

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).

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.

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	Number of stations north of 60° N	Number of stations north of 66° N
GLRIPD sites				
Total number of sites	465	317	244	90
Sites with post-1996 data	0	0	0	0
Sites with post-1990 data	152	110	88	44
Sites with post-1980 data	353	257	213	73
Sites with post-1970 data	381	273	222	76
Long-term sites (over 150 years) used by Magnuson et al. (2000)	9	6	6	1

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., 2009a; 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
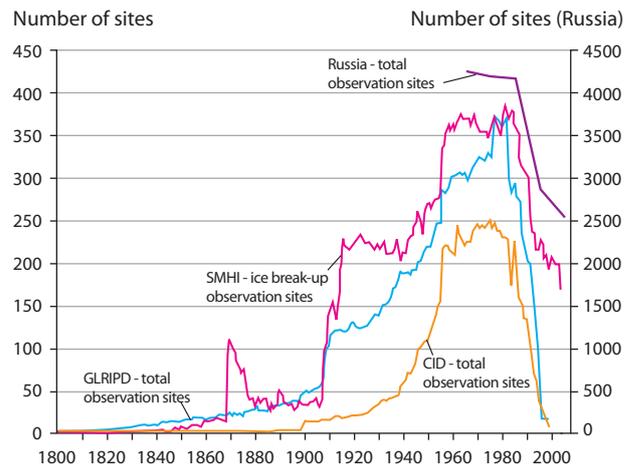
Figure 6.4. Historical evolution of the number of lake-ice and river-ice observation sites recorded in various databases. Data are from the Global Lake and River Ice Phenology Database (GLRIPD) (Benson and Magnuson, 2007); the Canadian Ice Database (CID) (Lenormand et al., 2002); the Swedish Meteorological and Hydrological Institute (SMHI, provided by G. Weyhenmeyer); and the Russian State Hydrological Institute (provided by V. Vuglinsky; note the scale difference for this latter data set).

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., 2010). 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
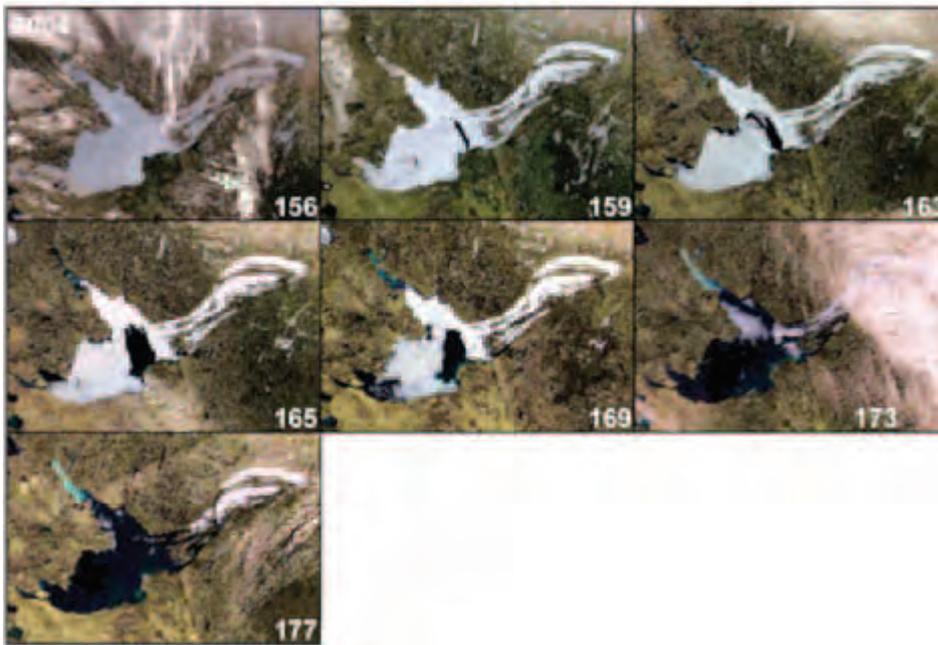
Figure 6.5. MODIS imagery from 2004 showing ice break-up on Great Slave Lake, with a crack and initial melt forming near the Slave River inflow. Numbers indicate Julian days. Source: Howell et al. (2009).

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
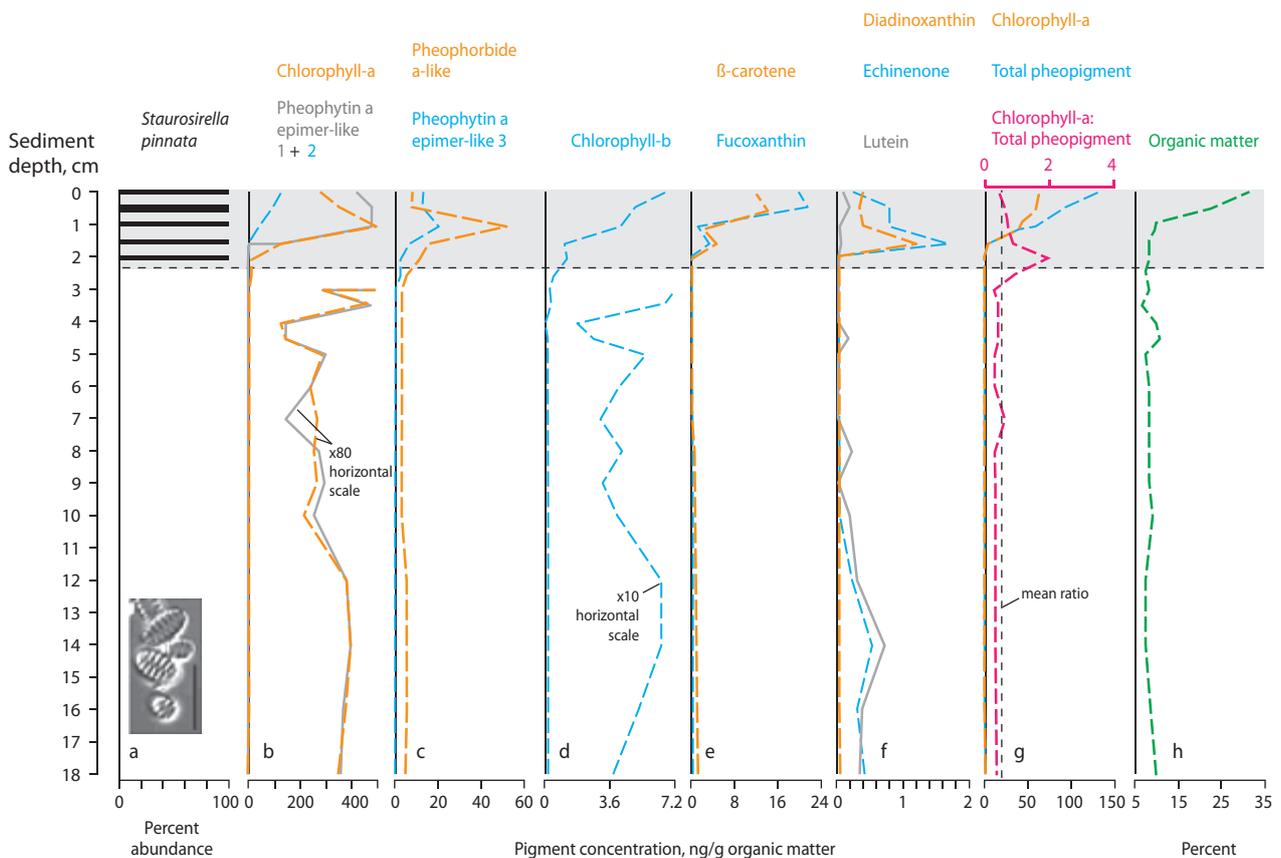
Figure 6.6. Change in the abundance and composition of algal communities over the past eight millennia, as inferred from fossil diatom and pigment profiles in Ward Hunt Lake sediments. Ward Hunt Lake is located on a small island north of Ellesmere Island in the Canadian Arctic. The most pronounced shifts occurred during the past two centuries. These abrupt changes are associated with a loss of ice cover and associated limnological and bio-optical changes in the water column. Recent change is indicated by the grey shaded area at the top of the figure. After Antoniades et al. (2007), which gives full details for each pigment profile.

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 were not 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 17.4 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 parts of Canada for the most rapid depletion period of 1970 to 2004. The degree to which this reflects the effects of either more
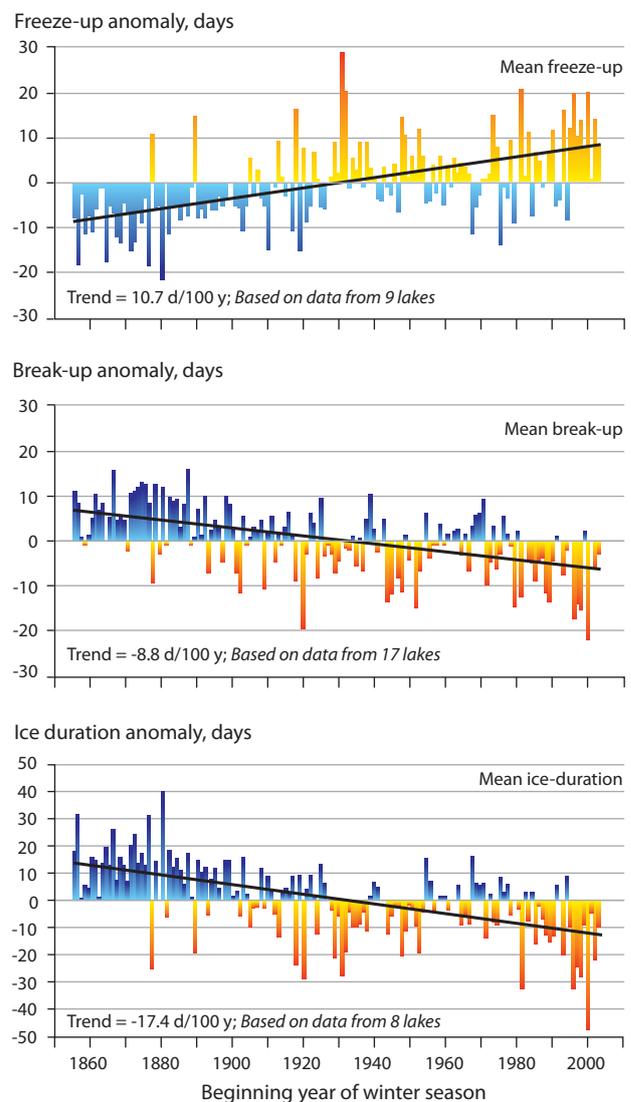
Figure 6.7. Long-term records of freeze-up, break-up and ice-cover duration for northern-hemisphere lakes. The plots present a 2007 update by B.J. Benson and J.J. Magnuson (Koç et al., 2009) of data originally presented by Magnuson et al. (2000) and based on records contained in the Global Lake and River Ice Phenology (GLRIP) database housed at the National Snow and Ice Data Center, University of Colorado at Boulder, USA. Note: stations are non-Arctic but included because of their unique long-term record length. After Koç et al. (2009).

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[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).

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. (2009b) 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. (2009b) 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

Maximum ice cover thickness, cm
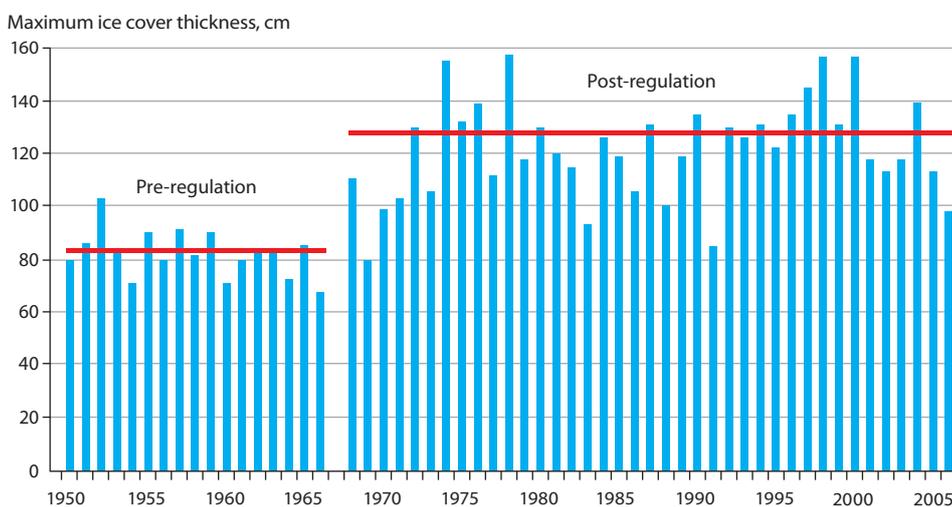
Figure 6.8. Step-change in maximum ice-cover thickness at the Tuoy-Haya station before regulation (river channel, 1951 to 1967) and after regulation (nearshore zone of Vilyuiskoye reservoir, 1969 to 2007). Data sourced from the State Hydrological Institute, St. Petersburg, Russia.

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
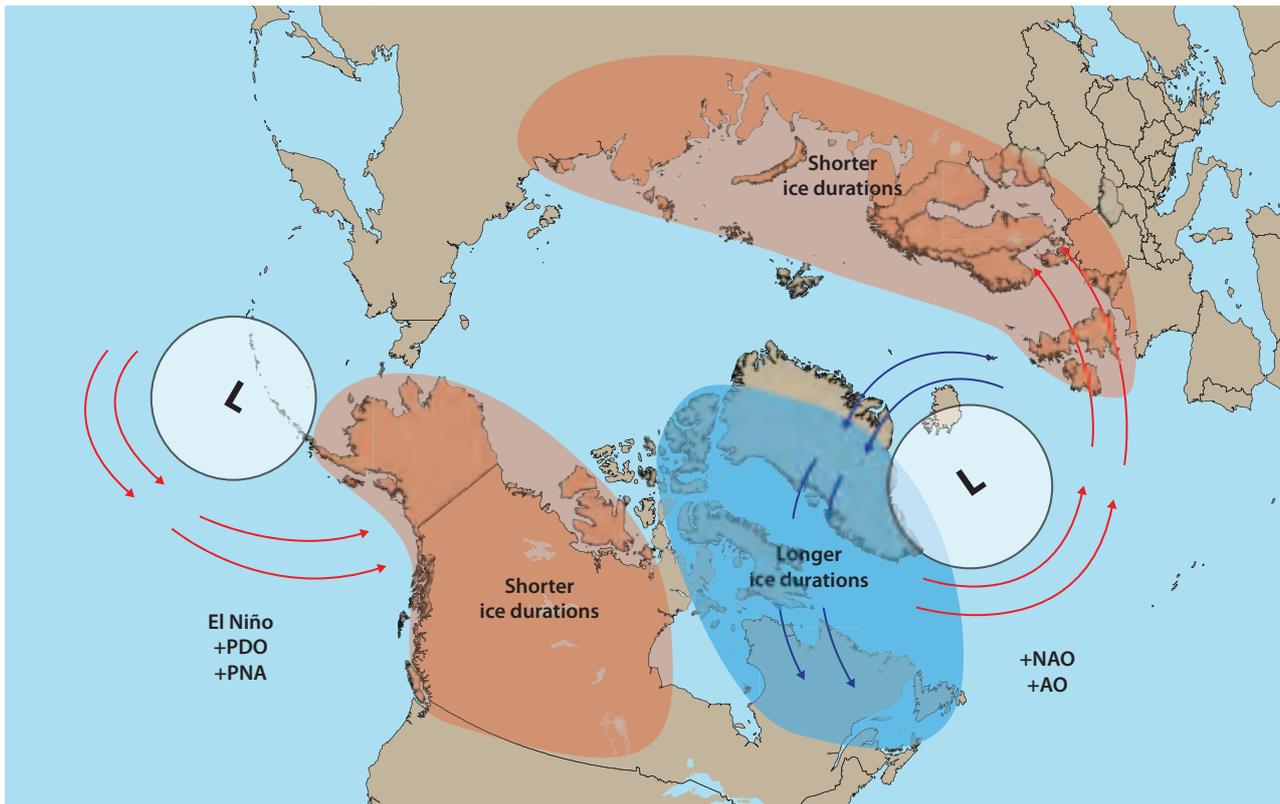
Figure 6.9. Schematic illustration showing typical cold-season atmospheric circulation patterns associated with the El Niño / Southern Oscillation (ENSO), the Pacific Decadal Oscillation (PDO), the Pacific North American (PNA), and the North Atlantic Oscillation / Arctic Oscillation (NAO/AO). Positive phases of ENSO, the PDO, and the PNA (shown here) are associated with a deeper than normal Aleutian Low causing higher temperatures and resultant shorter freshwater-ice duration over northwestern North America. Positive phases of the NAO/AO (shown here) are associated with a deeper than normal Icelandic Low causing lower temperatures and resultant longer freshwater-ice duration over northeastern Canada, and higher temperatures and resultant shorter freshwater-ice duration over Europe and northern Asia. Note that opposite phases of these circulation patterns (i.e., La Niña events and negative PDO, PNA, NAO/AO) are associated with weaker than normal Aleutian and Icelandic Lows, causing opposite temperature changes and freshwater ice durations than shown here.

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 global climate 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
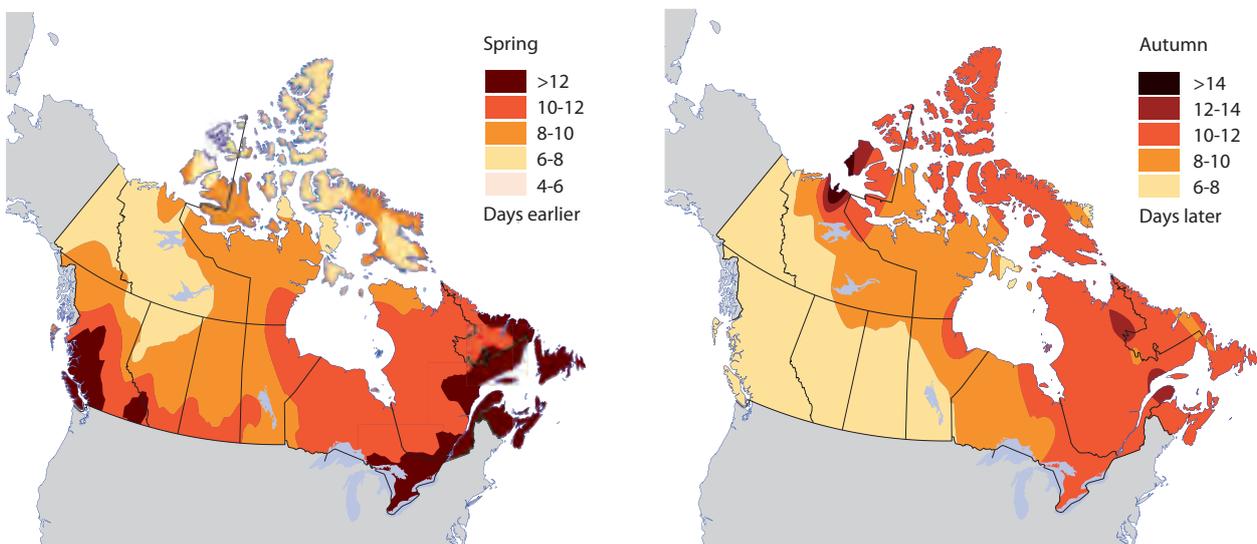
Figure 6.10. Projected changes to spring and autumn 0 °C isotherm dates over Canada for the 30-year period centered on the 2050s (2040–2069). Source: Prowse et al. (2007a).
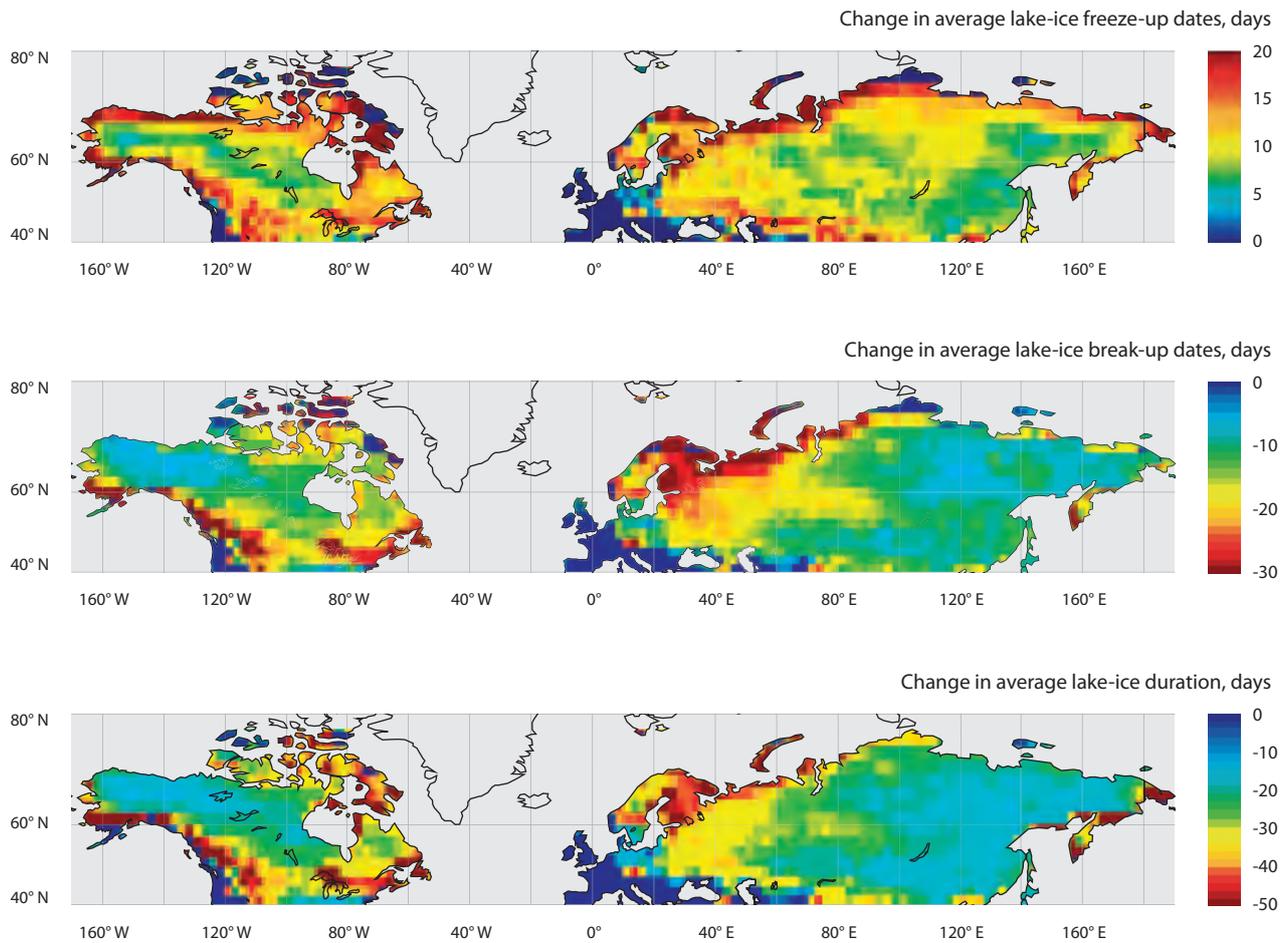
Figure 6.11. Change in simulated mean dates of freeze-up, ice break-up, and ice-cover duration between current (1960–1999) and future (2040–2079) climatic periods. Source: Dibike et al. (2010).

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
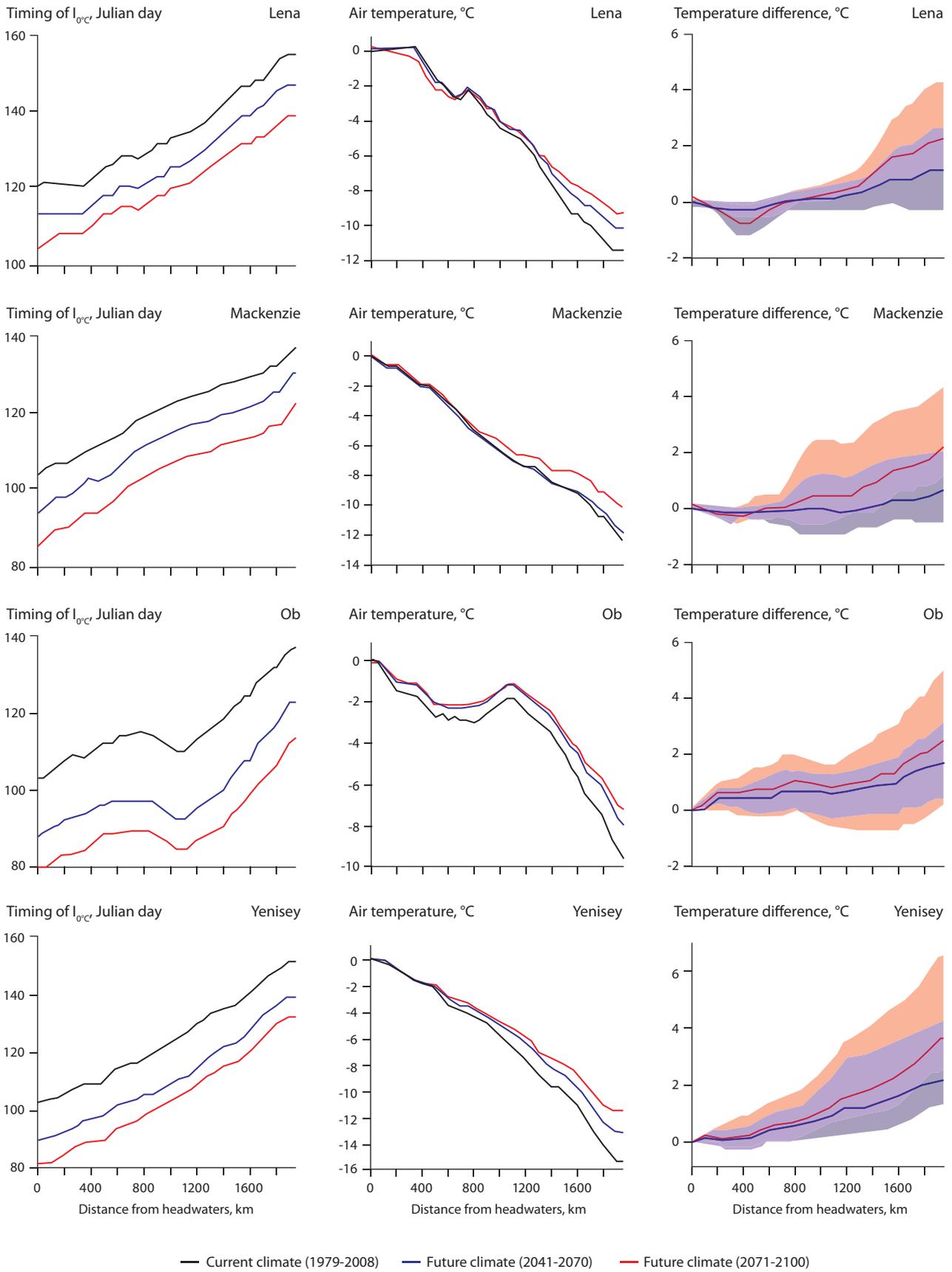
Figure 6.12. Projected changes in spring air temperature (average March to May) along the four largest Arctic rivers for two future time periods (2041–2070 and 2071–2100) compared to the current climate (1979–2008): (left column) current and projected 0 °C isotherm dates along the main-stem reaches of the four large Arctic rivers; (centre column) current and projected air temperature at the time when the upstream sections of the 2000-km main-stem reach 0 °C; and (right column) associated differences in air temperature between current and future climates. The solid blue and red lines are the GCM-ensemble means, while shading denotes the full range of projected GCM values. Source: Prowse et al. (2010).

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; Buzin and Kopaliani, 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-
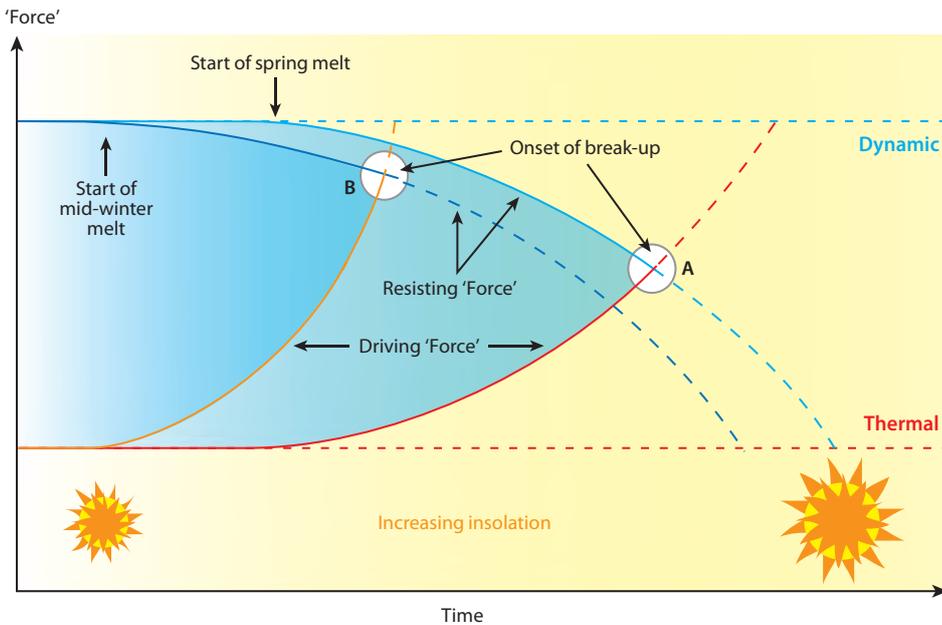
Figure 6.13. Schematic illustration of driving and resisting factors leading to different types of river-ice break-up events. Point A refers to a spring break-up and point B to a mid-winter event with smaller decreases in ice resistance because of lower levels of incoming solar radiation. Adapted from Beltaos (1982).

up events in Norway (Knut Alfredsen, Norwegian University of Science and Technology, Trondheim, 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 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) and Buzin and Kopaliani (2007, 2008). 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
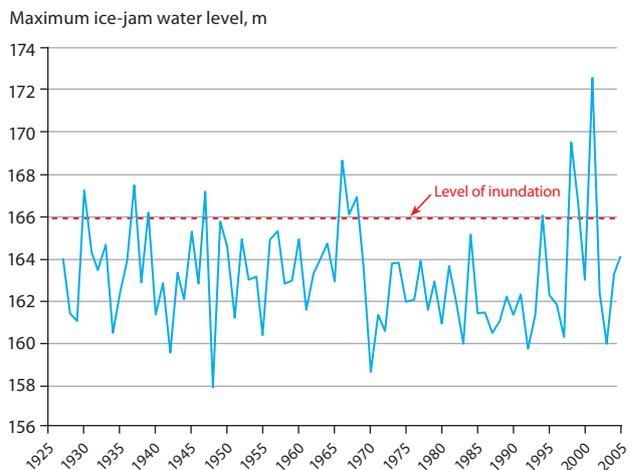
Figure 6.14. Chronological diagram of maximum ice-jam water levels on the Lena River near the city of Lensk, Russia. The vertical axis represents water level referenced to a local datum. The red line refers to the local level of flood inundation. Data provided by V. Vuglinsky, Russian State Hydrological Institute.

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 Section 11.2, this volume, on sea level change) 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
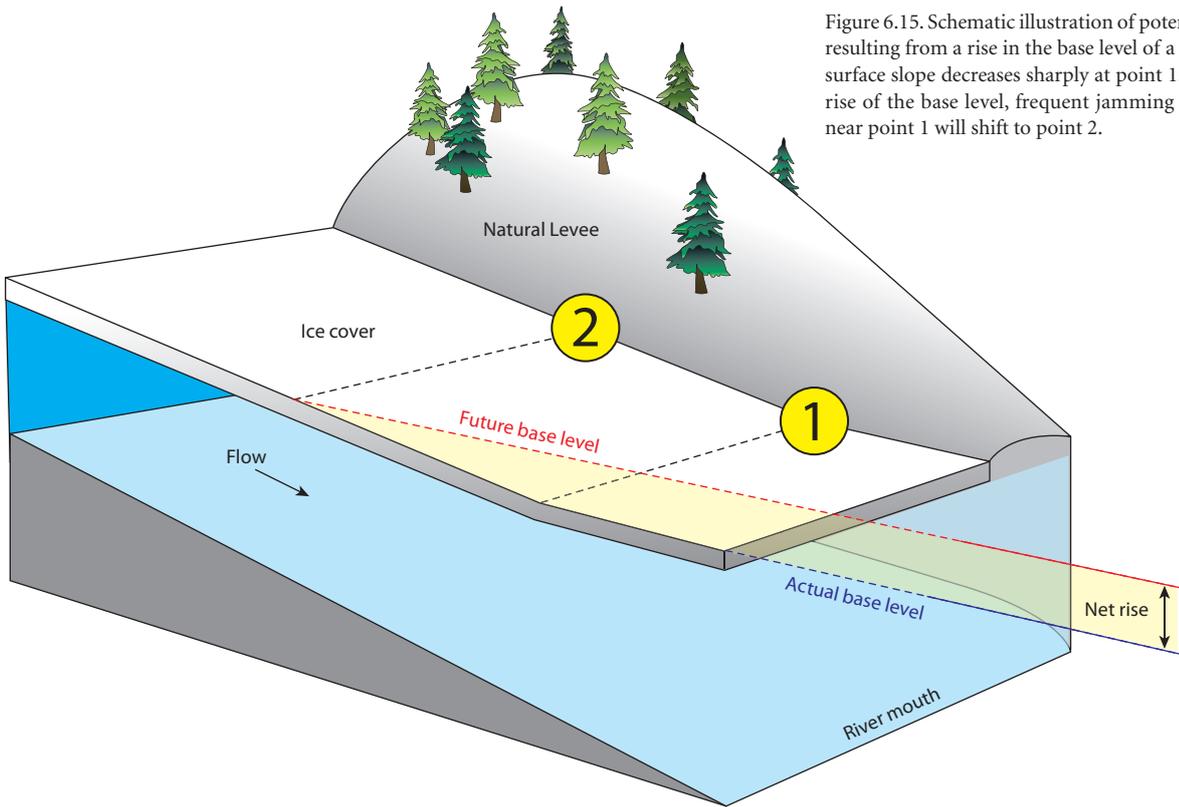
Figure 6.15. Schematic illustration of potential change resulting from a rise in the base level of a river. Water surface slope decreases sharply at point 1. With a net rise of the base level, frequent jamming that occurs near point 1 will shift to point 2.

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.

Discharge, m³/s
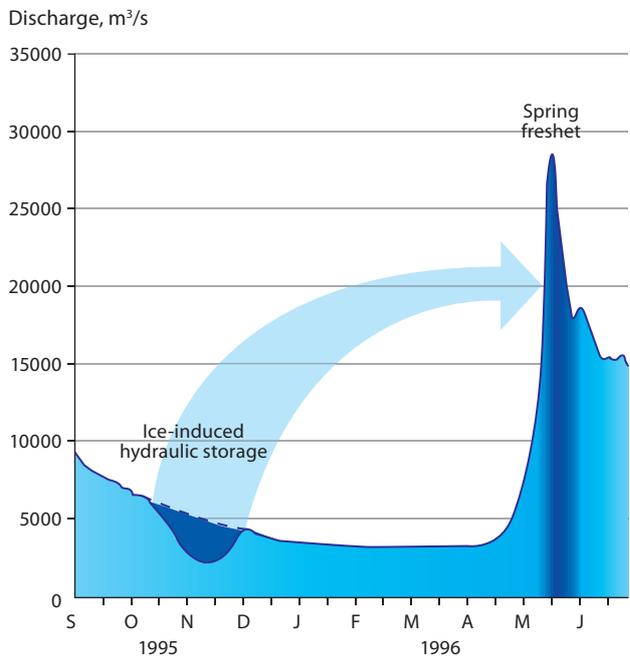
Figure 6.16. Occurrence of low-flow depression from ice-induced hydraulic storage and subsequent release during the spring freshet in the Mackenzie River, Canada. Storage occurs during autumn freeze-up (November), shown as the dark blue area of the declining hydrograph in about November to December. The line above the depression indicates assumed flow without this effect. Release of stored flow occurs during the spring hydrograph noted by dark blue shading. The magnitude of spring events is likely to decrease if ice-storage effects in seasonally transferring water are lost under future climatic conditions. After Prowse and Carter (2002).

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.
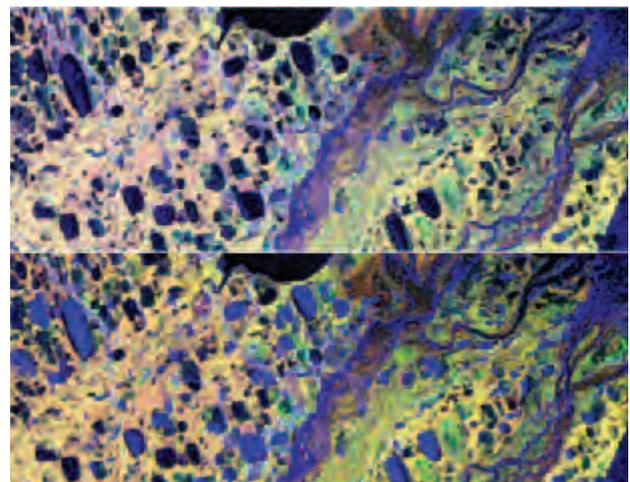
Figure 6.17. Images from ERS-1 (top: 20 April 1993) and ERS-2 (bottom: 16 April 2003) satellites showing ice cover over an area of the North Slope of Alaska (Sagavanirktok River). The light blue areas on the shallow lakes and river channels indicate the presence of ice with water underneath, while the dark blue and black areas represent lakes and river channels that are frozen to their bed (i.e., grounded ice). In 1993, the conditions were colder (thicker ice) than in 2003, which explains why more ice was frozen to the bed. Source: unpublished data but based on techniques outlined by Brown et al. (2010).

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., 2008a).

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 m² 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 may be attributable 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
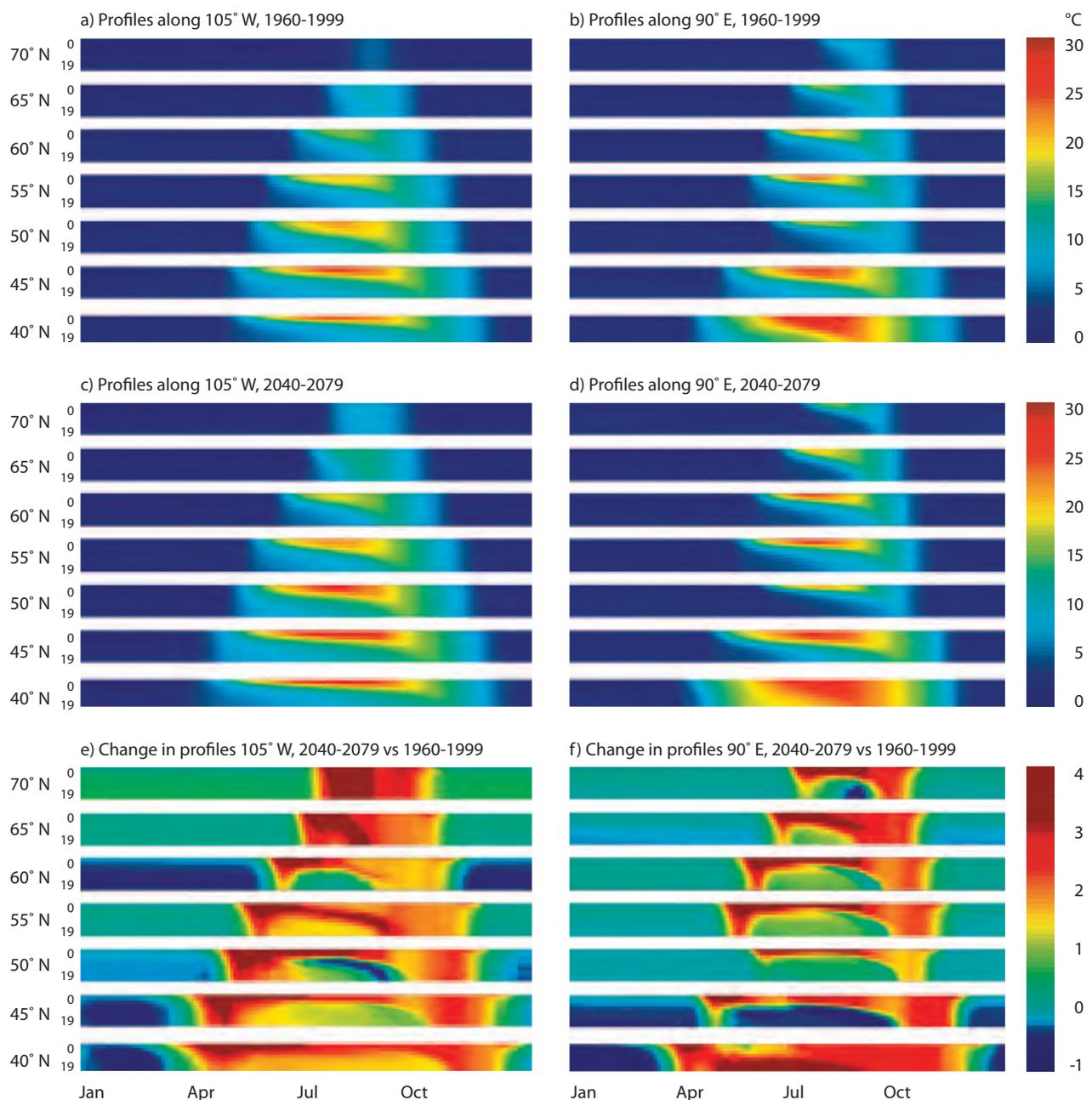

Figure 6.18. Modeled mean annual cycle of daily profiles in water temperature in lakes during (a and b) 1960–1999 and (c and d) 2040–2079 as well as (e and f) the corresponding change between these periods at 5° latitude intervals along example longitudinal transects on (a, c, and e) 105° W and (b, d, and f) 90° E. Source: Dibike et al. (2010).

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).

Figure 6.19. Methane bubbles trapped within lake ice cover (top: photo by Katey Walter Anthony as part of a NASA-funded research project) and ignition of methane when released and lighted (bottom: UAF photo by Todd Paris).

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 component 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, suggested 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., 2008b), 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., 2008a), 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.
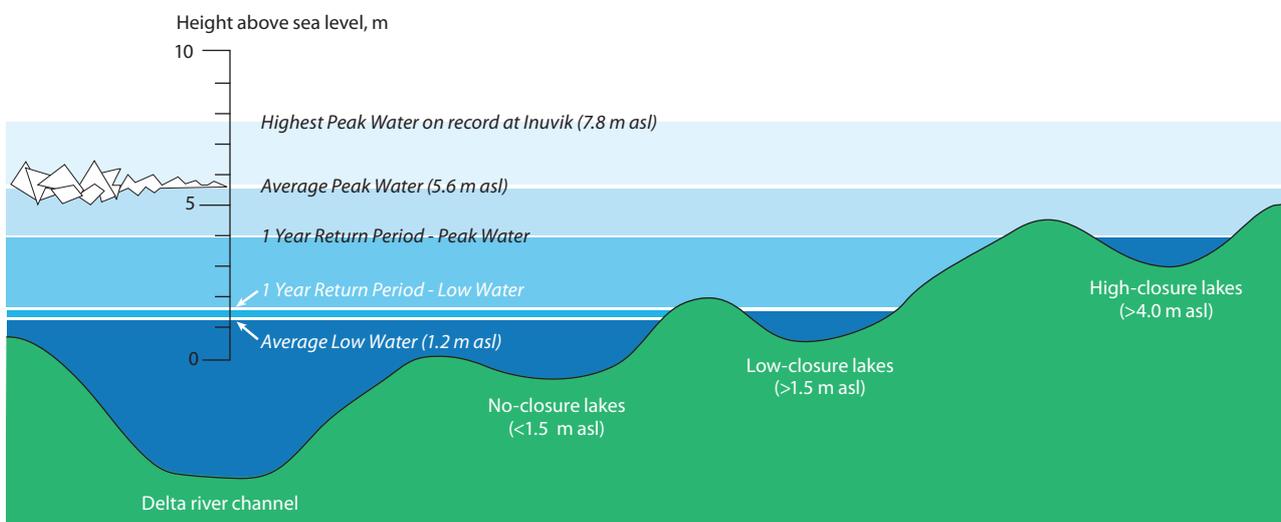

Figure 6.20. Classification of lakes in the Mackenzie River delta according to the extent of their isolation from the river. Ice-jam floodwaters are responsible for flooding of the riparian lake systems more hydraulically unconnected to the main flow system. Source: after Emmerton et al. (2007).
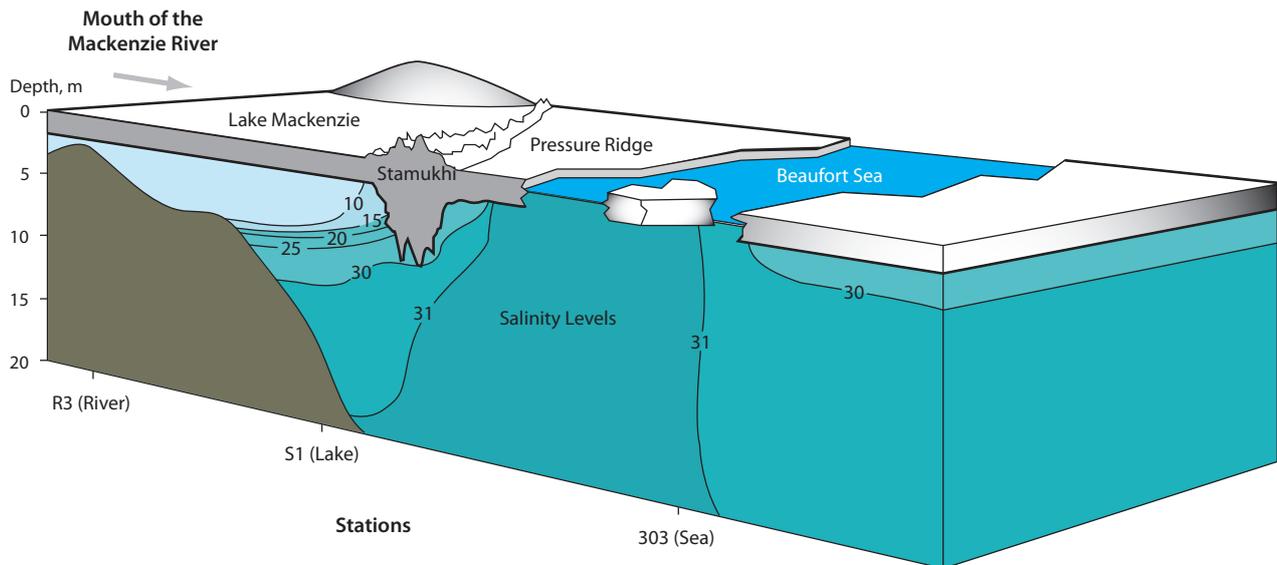
Figure 6.21. Schematic illustration of a stamukhi lake. The inflow from the Mackenzie River is trapped near shore behind the stamukhi (pressure ridge of sea ice), forming Lake Mackenzie. Source: modified from Galand et al. (2008a).

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 an 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
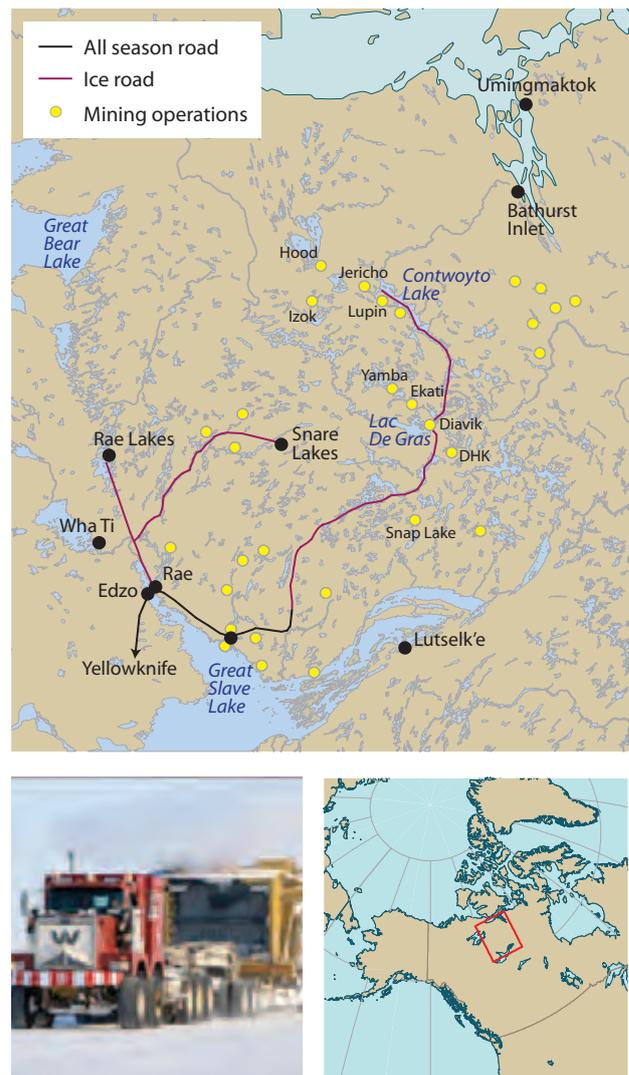

Figure 6.22. Example of large loads transported on the Tibbitt to Contwoyto Winter Road in northern Canada, which is composed of many long reaches across frozen lakes and rivers. Photo courtesy of Joint Venture Management Committee, partnership between DeBeers Canada, Inc.; BHP Billiton, Ltd; and Diavik Diamond Mines, Inc.

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
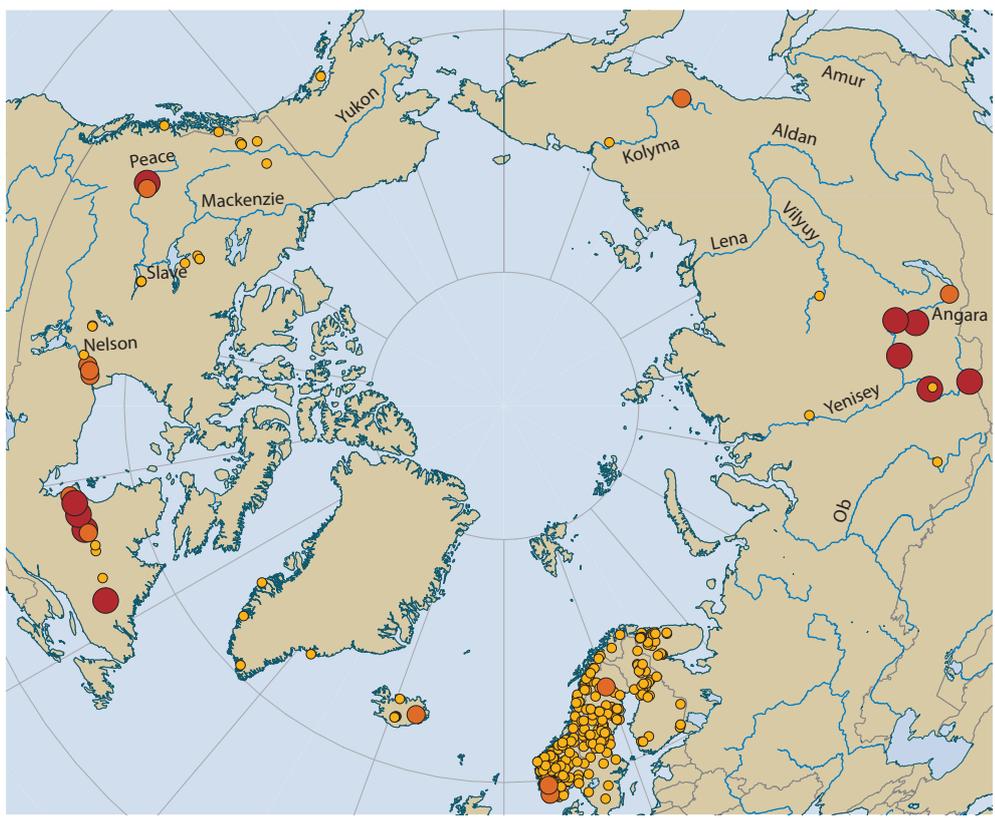
Figure 6.23. Location and size of major hydroelectric facilities located in ice-dominated river regimes of the circumpolar North. All such stations will require major modifications to current operational regimes and possibly infrastructure modifications based on projected changes to freshwater ice regimes (due to data availability, the figure may not show all existing plants). Source: Knut Alfredsen, Norwegian University of Science and Technology, Trondheim.

Installation, MW
· < 600
● 600 - 2000
● > 2000

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.

Acknowledgments

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