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Potential Impacts of Future Climate Change on Permafrost and Water Resources of Nunavut

BY:

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NWRI Contribution No: 05-306

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POTENTIAL IMPACTS OF FUTURE CLIMATE CHANGE ON PERMAFROST AND WATER RESOURCES OF NUNAVUT

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March 31, 2005

NWRI Cont. #05-306

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

PREFACE

Report Objectives

This document reports the potential impacts of future climate change on the permafrost and hydrology of Nunavut. Reports released by the Intergovernmental Panel on Climate Change (IPCC, 1996, 2001), and the Arctic Council and International Arctic Science Committee (IASC) (ACIA, 2005, 2004), have identified an accelerated advance of climate change over the latter half of the 20th century. Their projections of future climate change and associated impacts on the physical, biological and socio-economic nature of the Arctic over the 21st century indicate that climate change will not only continue, but will be most marked in the Arctic. This report examines the nature of climate change and the Arctic environment, the role of modeling in projection of future climate, the history of climate change in Nunavut and the Circumarctic, and the projected changes in climate, landscape and water resources of Nunavut as assessed by the ACIA (2005).

Literature Sources

This report is based primarily on the *Arctic Climate Impact Assessment* (ACIA, 2005), a project of the Arctic Council and the IASC. The ACIA assesses the impacts of projected climate changes on the Arctic over the 21st century, and is the first report to address regional variability of future climate change in the Circumarctic. Most of the material used in this report was drawn from Chapter 1 Introduction (Huntington *et al.*, 2005a), Chapter 2 Arctic Climate – Past and Present (McBean *et al.*, 2005), Chapter 3 The Changing Arctic: Indigenous Perspectives (Huntington *et al.*, 2005b), Chapter 4 Future Climate Change: Modeling and Scenarios for the Arctic Region (Källén *et al.*, 2005), Chapter 6 Cryospheric and Hydrologic Variability (Walsh *et al.*, 2005), Chapter 7 Arctic Tundra and Polar Desert Ecosystems (Callaghan *et al.*, 2005), Chapter 8 Freshwater Ecosystems and Fisheries (Wrona *et al.*, 2005), Chapter 9 Marine Systems (Loeng *et al.*, 2005), Chapter 15 Human Health (Berner *et al.*, 2005), and Chapter 16 Infrastructure: Buildings, Support Systems, and Industrial Facilities (Instanes *et al.*, 2005).

Also included in this report are:

- *Canada Country Study Volume II: Responding to Global Climate Change in Canada's Arctic* (Maxwell, 1997), which, based on the IPCC (1996) report on climate change, assessed climate change impacts and adaptations in physical systems, social environments and economics of the various regions of Canada;
- IPCC Climate Change 2001 reports (*Scientific Basis; Impacts, Adaptation, and Vulnerability: Chapter 16 Polar Regions; Synthesis Report*), which include a comprehensive review of climate change and the science of projecting future climates, and identification of future impacts of climate change globally, along with adaptations and vulnerabilities;
- Canadian Council of Ministers of the Environment (CCME, 2003) report on *Indicators of Canada's Changing Climate*, which reviewed recent and ongoing changes to Canada's climate and the effects of these changes; and,
- up to date scientific publications that address climate change in Nunavut.

<u>Report Structure</u>

This report is structured to provide the reader with an introductory understanding of the climate system and climate variability, and a more detailed description of the projected impacts of climate change on the landscape and freshwater resources of Nunavut. The first part of the report describes the climate system and the nature of climatic change. This is followed by a discussion of the vulnerability of the Arctic to change. A historical account of climate change and associated impacts is then presented. The report concludes with a description of projected changes in temperature and precipitation for Nunavut and the Circumarctic, and discusses the potential impacts of these changes on sea ice, glaciers and ice sheets, snow, permafrost, and river and lake ice, as well as the effects of these changes on water quality and infrastructure.

Reading this Report

A number of considerations must be taken into account when reading this report. First, this report is based upon a compilation of literature pertaining to climate change and effects in the Arctic. Though literature addressing climate change in Nunavut does exist, the majority of the conclusions regarding future climate change and impacts on the territory of Nunavut are inferred from international studies and assessments of climate change impacts at global and coarse regional scales (ACIA, 2005; IPCC, 2001). Where possible, reference to Nunavut-specific documentation has been made, and issues specific to the physical and socio-economic environments of the territory have been addressed. For the purposes of this report, the ACIA-projected changes in temperature, precipitation, degree-day and potential evaporation for the Arctic through the 21st century were recalculated for territory of Nunavut (Bill Chapman, unpublished data).

Second, projected changes in temperature and precipitation are derived from highly variable output generated by General Circulation Models (GCMs). The projected changes in future climate and its effects are, therefore, based upon an average from a number of models (e.g., 5 coupled Atmosphere-Ocean GCMs used in the ACIA). Model-averaged climate simulations have been shown to quite accurately reproduce past climate. Climate reconstructions for the latter half of the 20th century, however, have been somewhat less successful than those for earlier periods (Weatherhead *et al.*, 2002). IPCC (2001) and ACIA (2005) model-averaged projections of future changes in temperature and precipitation are, however, similar, increasing confidence in these models (see Källén *et al.* (2005) for a detailed discussion of the techniques, strengths and limitations of climate modeling).

Third, the certainty with which projections of climate change impacts on high latitudes may be made, is restricted due to limited research and monitoring in the Arctic. High-latitude research and instrumental records, hence knowledge of climate-environment interactions, are limited relative to more temperate locations. Assessed impacts of climate change on Nunavut are, therefore, based upon a) knowledge of the general structure and function of marine, freshwater, terrestrial, etc. systems and their potential responses to climate change, as recorded in limited historical archives (e.g., ice cap and lake sediment cores) and abbreviated instrumental climate records, b) climate change effects as recorded in other regions of the Arctic and more temperate locations, c) recent observations of climate change impacts, and d) consideration of regional and local features and their potential response to, and interactions with, climate. The certainty of future responses to climate change is also complicated by the complex nature of climate-landscape-water interactions, and the potential for multiple and varying effects in different systems in various locations (Wrona *et al.*, 2005). As well, the effects of climate change may be expressed in a threshold, non-linear, synergistic or cumulative manner, which may positively or negatively feed back to climate change.

EXECUTIVE SUMMARY

Introduction

Global climate of the 20th century has changed at a rate unprecedented in the last thousand years. The greatest changes in climate and some of the most dramatic effects of climate change have occurred in the Arctic, and are projected to be most pronounced at high latitudes as climate change progresses through the 21st century. The Arctic Climate Impact Assessment (ACIA) was an international, circumarctic initiative to assess the regional impacts of projected climate change and UV radiation on the physical, biological, socio-economic and cultural environments of the Arctic over the 21st century. The current report examines the potential changes in permafrost and water resources of Nunavut with future climate as projected by the ACIA.

Arctic climate system

The Earth's climate is a manifestation of interactions and feedbacks between the atmosphere, hydrosphere, cryosphere, land surface and the biosphere, as well as humans. As such, Earth's climate varies in response to both natural and anthropogenic forcing. The cryosphere exerts significant control on the climate of the Arctic, reflecting radiant energy back to the atmosphere and positively feeding back to cooling of terrestrial and aquatic surfaces in the Arctic. Temperature and precipitation, in turn, affect the thermal state of the cryosphere, which is highly responsive to changes in temperature relative to surfaces in more temperate environments. As such, the Arctic environment and the people that are adapted to, and rely upon, traditional resources are highly vulnerable to future climate change.

Historical climate change and effects

Global and arctic climate are naturally variable. The high variability of 20th century climate and the dramatic rise in air temperatures over the latter part of the century, however, caused substantive changes in terrestrial and aquatic systems of the Canadian Arctic. Sea ice generally declined in thickness and extent. Some glaciers and ice caps of Nunavut lost mass, while others grew in response to increased precipitation. Though snowfall generally increased in arctic Canada, snow cover extent declined due to warmer and earlier spring melt and, in some instances, later fall cooling. Milder climate and abbreviation of the snow- and ice-covered seasons contributed to increased freshwater discharges in some rivers of Nunavut, decreased discharges in others, and longer open-water season in freshwaters of the Arctic Archipelago. Warmer temperatures and longer ice-free season resulted in drying of some lake and rivers in Nunavut. Recent warming also contributed to rising permafrost temperatures, which, in some locations of the Arctic, altered stability and moisture regimes of land surfaces. Although climate change and associated effects over the latter part of the 20th century have varied highly regionally and locally, the general trend has been toward warming, increased precipitation, and degradation of the cryosphere.

Future climate change and effects

The ACIA modeled holarctic and regional changes in temperature and precipitation, as well as various other climate parameters, and assessed the impacts of these changes on the physical landscape, the biological nature, and the socio-economic and cultural aspects of the Arctic. Future changes in climate and associated effects are projected to be heterogeneous, varying regionally and locally, as well as temporally, as climate change progresses. Overall, from 1981-2000 to 2071-2090 the average annual air temperature is projected to increase by 3.7°C, precipitation by 12.3%, and evaporation by 15.5%. Average warm and cold season air temperatures for Nunavut are projected to increase by 2.6 and 4.7°C, respectively, with precipitation increasing by 21.4 (warm) and 15.1 mm (cold), degree-days by 60.5 (warm) and 7.8 days (cold), and potential evaporation by 4 (warm) and 0.9 mm (cold). These results indicate that, generally, Nunavut will likely be warmer and wetter in the future, with the most pronounced increases in air temperatures occurring over the fall and winter, and at higher latitudes. Though precipitation will likely increase in the future, increasing length of the warm season and increased evaporation may offset this effect.

Altered seasonal and spatial temperature and precipitation regimes will affect the land and water resources of Nunavut and the circumarctic by the end of the 21st century. Sea ice extent is projected to decline by between 12 and 46%. Glaciers and ice sheets are expected to melt, contributing to between 0 and 6 cm rise in sea level. Snow cover is projected to decline by 9 to 17%, with the greatest reductions occurring in spring and winter, and with snow distribution shifting northward. Circumarctic permafrost is projected to decrease in area by 23%, with continuous permafrost declining in area by 41% by the end of the 21st century, and distribution of the discontinuous permafrost zone shifting northward. Active layer depth is expected to increase by 30 to 50%. Though continuous permafrost of Nunavut may not change substantially in distribution to 2080, particularly beyond the southern boundary, the active layer is projected to increase by greater than 50% by 2050. Freshwater ice is projected to decline in thickness and duration, with earlier melt in spring and later freeze in fall.

As these changes to freshwaters and permafrost progress over the 21st century, they will be accompanied by changes to the quality of water resources and the integrity of infrastructure. Water quality may be expected to degrade with future climate change due to increased loading of contaminants, sediments and nutrients from snow, ice and permafrost melt, increased atmospheric deposition, increased human activity, and damage to sanitation and water distribution infrastructure. The stability and life of present-day infrastructure will be threatened by permafrost melt and increased frequency and magnitude of extreme hydrologic events.

I. INTRODUCTION

The Earth's climate warmed over the 20th century at a rate unprecedented over the preceding millennium, and is projected to continue warming in an accelerated manner through the 21st century (ACIA, 2005; IPCC, 2001a). The magnitude of climate warming, hence the severity of its impacts, has been, and is expected to continue to be, greatest in the Arctic. The extreme nature of arctic landscapes and environments make the Arctic, and the people that depend on its resources, particularly vulnerable to the effects of climate change. The impacts of future climate change will necessitate human responses that integrate knowledge of the changing environment, adaptive practices, and sustainable management of resources.

Global climate models predict that over the 21st century temperatures in the Arctic will rise dramatically, precipitation will increase, and the growing season will lengthen. This report reviews the impacts of these changes on the landscape and water resources of Nunavut, drawing heavily upon the Arctic Climate Impact Assessment (ACIA, 2005). The current report begins with a brief review of the functioning and variability of the Earth's climate system, a discussion of the interactions between the various components of the climate system in the Arctic, and an examination of the sensitivity of the Arctic to climate change. An overview of historical climate change and associated impacts in the Arctic is presented. This is followed by a description of projected changes in temperature and precipitation in Nunavut and the Circumarctic over the 21st century, and a discussion of the potential impacts of these changes on sea ice, glaciers, snow, permafrost and ice, water quality, and infrastructure in Nunavut.

II. EARTH'S CLIMATE AND CHANGE

Climate may be defined as the average weather and its variability over a defined time and space (IPCC, 2001). The Earth's climate is naturally variable over a variety of temporal and spatial scales. This variability is a manifestation of external forcings on the climate system and internal forcings or interactions between climate system components (atmosphere, hydrosphere, cryosphere, land surface, biosphere). The discussion below is based on the overview of the nature and functioning of the global climate system as presented in Baede *et al.* (2001).

Earth's climate is generated by an exchange of energy between the sun (incoming solar radiation) and the Earth's atmosphere (outgoing infrared radiation). This exchange of energy is affected by a number of factors including the composition of the atmosphere (e.g., gases, aerosols), cloud cover, and surface temperature and albedo (i.e., proportion of incident light reflected back to the atmosphere). Imbalances in this exchange of incoming and outgoing radiation cause changes in the Earth's climate and are generated by external forcing (e.g., fluctuation in incoming solar radiation with orbital variability) and internal forcing (e.g., reflection of incoming radiation by high albedo snow and ice surfaces, hence cooling of the Earth's surface). Positive radiative forcing results in net warming on Earth, and negative forcing produces net cooling.

Variability in radiative forcing arises not only from natural fluctuations in solar energy and interactions between climate system components, but from human activities as well. Although land use activities have changed the albedo of surfaces in many locations (e.g., urban surfaces typically have lower albedo than do more natural, vegetated surfaces), hence the temperature of the Earth's surface and associated radiative forcing, the most substantial effect that human's have had on the climate system has been alteration of the Earth's atmosphere over the last 150 years. Since the middle of the 18th century human activities (e.g., fossil fuel, biomass burning) have expanded at an exponential rate, and atmospheric concentrations of greenhouse gases and aerosols have risen significantly relative to the "stable" levels that persisted through the preceding millennium. Atmospheric concentrations of carbon dioxide continue to rise at an unprecedented rate of 0.4%y^{-1}. Isotopic analyses indicate that these excesses of carbon dioxide originate from anthropogenic sources. Concentrations of other greenhouse gases (e.g., methane, nitrous oxide, tropospheric ozone) have risen as well over the 20th century, as have atmospheric concentrations of aerosols (e.g., dust, sulphates, nitrates). Greenhouse gases trap energy at the Earth's atmosphere, reducing emission of radiation back to space. Rising atmospheric concentrations of these gases, therefore, generate positive radiative forcing (warming).

III. ARCTIC LANDSCAPE, WATER AND PEOPLE: SENSITIVITY TO CLIMATE CHANGE

The IPCC (2001b) concluded that climate change effects will be most pronounced in the Arctic. Snow, ice and frozen soils are intrinsic to high latitudes, shaping land and water resources, socioeconomics and culture (Arctic Human Development Report, 2004). This prominence of the cryosphere contributes highly to the sensitivity of the Arctic to climate change. The vulnerability of the Arctic to change varies regionally and locally, with potential for dramatic climate impacts in some areas.

The cryosphere is highly sensitive to changes in temperature (Anisimov *et al.*, 2001). Although frozen soils, snow, glaciers and ice positively feed back to cooling at high latitudes (e.g., reflection of radiant energy off snow and ice cools the Earth's surface, hence limiting melt of the cryosphere), even slight increases in temperature can cause dramatic responses, fundamentally altering aquatic and terrestrial systems. The effects of climate change may, therefore, be expected to be magnified in the Arctic relative to more temperate regions.

This sensitivity of Arctic environments to temperature change may contribute to cascading effects that further amplify climate change impacts. Changes to climate in the Arctic will not only have direct effects, but may also produce feedback effects (Everett and Fitzharris, 1998). For example:

- melt of snow and ice increases exposure of ground and vegetation to radiation, thereby contributing to rising surface temperatures, which positively feed back to melt;
- melt of ice covers increases incident radiation at the water surface and raises water temperatures, which negatively feeds back to ice formation;

- warmer soil temperatures and longer-duration warm season increase soil microbial activity and associated generation of greenhouse gases, which, upon release to the atmosphere, positively feed back to climate warming through the greenhouse effect.

Threshold effects are of particular concern in the Arctic as there are important environmental thresholds associated with water-phase change (Anisimov *et al.*, 2001). As such, future responses of the Arctic to climate change may be dramatic and unpredictable when thresholds are exceeded (Wrona *et al.*, 2005). For example, permafrost is estimated to begin gradually thawing when average annual temperatures are above -2°C (J. Brown, personal communication in Wrona *et al*, 2005), and some lakes are susceptible to catastrophic drainage with permafrost thaw (Marsh and Neumann, 2001, 2003).

The sensitivity of arctic land and water resources to climate change varies circumarctically. For example:

- warm, thin permafrost, such as that in the discontinuous permafrost zone, is more sensitive to changes in temperature than is cold, thick permafrost at higher latitudes;
- high arctic environments, which have low annual precipitation and a very abbreviated summer, are more thermo-hydrologically sensitive to changes in heat and moisture than are lower latitude environments (Young and Woo, 2003);
- high arctic coastal areas have very thick sea ice and are, therefore, less vulnerable to melt and associated effects than are areas with thinner sea ice covers (Maxwell, 1997);
- coasts consisting of unlithified materials are more resistant to erosion by sea ice and permafrost melt than are those consisting primarily of bedrock;
- glaciers that are exposed to colder temperatures and more continental climate are less sensitive to changes in temperature and precipitation, hence more effective at moderating the effects of climate change than are warmer, more maritime glaciers (Walsh *et al.*, 2005); and,
- cold, nutrient-poor waters of arctic lakes and ponds are susceptible to significant changes in sediment, nutrient, carbon and contaminant loadings with shifts in

catchment permafrost, hydrology and vegetation, relative to those at more temperate latitudes.

In some instances, the impacts of climate change may be exacerbated by compounded effects. For example:

- coastal landscapes and water resources are susceptible not only to changes in terrestrial and freshwater systems, but to those in marine systems as well;
- large, freshwater river systems that flow from more temperate headwaters northward to the Arctic Ocean are vulnerable to climate change effects at high latitudes, as well as those at lower latitudes;
- the Arctic is vulnerable not only to the direct effects of climate change on aquatic and terrestrial systems, but also to potentially dramatic changes in land use; and,
- residents of the Arctic are highly dependent upon climate-sensitive resources, hence, particularly in the case of smaller, more traditional communities (e.g., Canada's eastern Arctic), are very vulnerable not only to future climate change effects on the physical environment, but effects on their health and culture as well (Hardy and Bradley, 1997).

IV. HISTORICAL CLIMATE CHANGE IN THE ARCTIC

Historical climate variability, over millennia and over the recent past, is an important tool in the assessment of potential future impacts of climate change. Records of paleoclimatic changes and impacts on terrestrial and freshwater systems may be archived in ice caps, permafrost, and lake sediments. More recent climate variations are recorded in instrumental measures, traditional ecological knowledge, and research. McBean *et al.* (2005) and Wrona *et al.* (2005) review tools for assessing past climate change, and examine changes in climate from approximately 120 million years ago to present day. Hardy and Bradley (1997) review historical changes in the climate of Nunavut.

Tools for the Study of Past Climate

Changes in climate are manifested in the physical, chemical and biological features of landscapes and ecosystems. Past changes to climate may, therefore, be investigated through examination of terrestrial and aquatic archives, instrumental observations, and observations by local residents. Climate reconstructions may be achieved with relative certainty using multiple climate proxies.

Climate reconstructions based on paleoclimatic proxy data are key to understanding the nature and potential effects of climate change, and the nature and direction of responses of hydrological, ecological and landscape systems to those changes. Ecosystem records are, therefore, invaluable to climate reconstruction, and are commonly archived in ice caps, soils and permafrost, and sediments. Proxies for climatic change include pollen, diatoms, sediment structures, and isotopic signatures, which can reveal information regarding growing season, growth rates, species and trophic structure, limnology, water levels, nutrients, air masses, precipitation, melt, erosion and runoff, treeline, and fire (Lamoureux and Gilbert, 2004; Gibson *et al.*, 2002; Rühland and Smol, 2002; Wolfe *et al.*, 2001; Moser *et al.*, 2000; Smol and Cummings, 2000; Douglas and Smol, 1999). Climate reconstructions from past millennia are often variable due to limited data and inconsistencies between proxy indicators. Nevertheless, multiple paleoclimatic proxies may be used to describe regional histories and to validate key conclusions (McBean *et al.*, 2005; Wrona *et al.*, 2005).

Climatic reconstructions of the past 150 years are generally of better quality and resolution than those of past millennia. Instrumental measures, as well as paleoclimatic data, experimental research, and a wealth of traditional ecological knowledge substantiate and bolster conclusions regarding past climates and environments (IPCC, 2001). These reconstructions are, however, limited by sparsely and unevenly distributed monitoring and research in the Arctic, and limited knowledge and understanding of climate-landscape-water resource interactions and feedbacks. Long-term instrumental records are less common in Nunavut than in more developed areas of the Arctic (Hardy and Bradley, 1997). Coupled paleoclimatic-instrumental reconstructions have been used in validating theories of anthropogenic forcing of the climate system, and in confirming observations of recent accelerated climate change (McBean *et al.*, 2005; Hanssen-Bauer and Førland, 1998).

Indigenous knowledge of landscape, ecology and resources is also very important to reconstruction and assessment of climate change and its impacts. For a review of North American indigenous observations of climate change in the Arctic please see Krupnik and Jolly (2002) and Huntington *et al.* (2005b). Traditional ecological knowledge assists in validation of scientific theories and claims regarding the nature of climate change and its impacts. For example, both scientific and traditional observations document advance in the breakup of river ice over the past 150 years (Huntington *et al.*, 2005b; Magnuson *et al.*, 2000). On the other hand, observations of increased shrub and lichen growth in the Kitikmeot region of Nunavut (Thorpe *et al.*, 2001) are contrary to experimental evidence that suggests a decline in new lichen growth in the region (Huntington *et al.*, 2005b). Though traditional knowledge and science may not always be consistent, both are essential to effective reconstruction and assessment of climate change.

Climate Change Prior to the 20th Century

Climate in the Arctic has alternated between numerous warm and cold periods over the last 2 million years. Records indicate that prior to the climatically variable Quaternary Period (1.6 million yBP to present), the Arctic was significantly warmer than in the 20th century. Quaternary climatic variations, manifested as glacial and interglacial stages with episodic stadials and interstadials (brief cool/mild periods; see Table 1), are theorized to have been driven by Earth's orbital variations (i.e., receipt of solar radiation, McBean *et al.*, 2005; Macdonald *et al.*, 2000). Resultant changes in sea surface temperatures and circulation patterns of oceans and the atmosphere, as well as changes to the hydrologic cycle, vegetation and land-ice cover, are theorized as having contributed to periods of glacial advance and retreat.

Table 1. Climate change prior to the 20th century (compiled from e.g., McBean *et al.*, 2005).

Years BP (approximate)	Details	Source
120-90 million	Significantly warmer than present	
Over last 2 million	Glacial and interglacial episodes	
2.54 million	Pre-Quaternary onset of NE North American glaciation due to orbital variations	Maslin *et al.* (1998)
1.6 million to present Quaternary	Periodic climatic variations i.e., glacial and interglacial, stadials and interstadials; driven by Earth orbital variations	Imbrie and Imbrie (1979), Berger (1988), Imbrie *et al.* (1993a,b)
800,000 to present	~10 glacial stages	Ruddiman *et al.* (1989), Ruddiman and Kutzbach (1990)
130,000 to ~117,000 Last Interglacial	Warmer (2°+) and drier than present interglacial in North America	IPCC (2001) Bourgeois *et al.* (2000)
~20,000 Last Glacial Maximum	The rate of temperature recovery from this point is a benchmark for assessing current rates of warming, e.g., average 2°+ per millennium between 20,000- 10,000 years BP in Greenland; lower elsewhere	
Retreat of Glacial Ice 13,000-8,500	Retreat of Laurentide Ice Sheet from W and central mid Arctic Retreat of fjord-based ice (Ellesmere Island) as late as 5500 Interrupted by cooling of Younger Dryas Interstadial	Dyke and Prest (1987) England *et al.* (2000)
11,000-9,000 Early Holocene	Climate variations due to orbital forcing Warmer than 20th century e.g., abrupt warming of central Greenland of 7°+ in a few decades Some areas not warm until after 8000 years BP because of Ice Sheet e.g., eastern North America Increased precipitation in Baffin Bay due to warm sea surface, causing glacial advance Extensive glacier cover in high Arctic.	e.g., Stewart (1988) Johnsen *et al.* (1992), Grootes *et al.* (1993), Severinghaus *et al.* (1998) Webb *et al.* (1998) Miller and Vernal (1992)
9,000-3,000 Mid Holocene	Warm e.g., Baffin warming 8000 - 3000 with max near 6000 years BP	Williams *et al.* (1995) e.g., Smith (2002)
1,000 AD to 1850-1900 Late Holocene	Modest and irregular cooling - Little Ice Age (1400-1800) Expansion of glaciers and formation of ice shelves e.g., off Ellesmere Island Coldest temperatures 100 to 299 years before 2000.	e.g., Jones *et al.* (1998), Mann *et al.* (1998,1999), Crowley and Lowery (2000) Blake (1989), Stewart and England (1983) Bourgeois *et al.* (2000)

The last glacial maximum occurred approximately 20,000 years BP, ending with a gradual, progressive recovery of warmer temperatures into the Holocene (approximately 11,000 yBP to present). Warming temperatures, likely resulting from a change in Earth's orbit (i.e., proximity to the sun), resulted in deterioration of the Laurentide Ice Sheet from the western and central mid Arctic (Dyke and Prest, 1987). Glacial retreat was delayed in eastern North America, with remnants of glacial ice persisting as late as 5,500 on Ellesmere Island (England *et al.*, 2000; Gajewski *et al.*, 2000).

Early and mid Holocene conditions, though regionally variable (e.g., areas of ice sheet persistence), were generally warmer than those of the 20th century (Gajewski *et al.*, 2000). Orbital variations during the early Holocene generated higher summer insolation and lower winter insolation north of 60°N (Kutzbach *et al.*, 1993), producing summer temperatures 1-2°C warmer than those in the late 20th century (Overpeck *et al.*, 1997). Ice masses in Nunavut deteriorated rapidly in response to this warming (e.g., Evans and England, 1992). Some areas of the Arctic warmed at a slower rate than others (e.g., eastern North America, Webb *et al.*, 1998) due to delayed retreat of glacial ice. Glacier cover in the high Arctic remained extensive, and increased precipitation over Baffin Bay contributed to glacier advance during the early Holocene (Miller and Vernal, 1992). Moisture availability was higher over the mid Holocene (Gajewski *et al.*, 2000), though regionally variable (Wrona *et al.*, 2005; Overpeck *et al.*, 1997).

Warming over the early and mid Holocene degraded into modest and irregular cooling over northern North America from 1,000 AD to 1850-1900. Summer temperatures were typically cooler, contributing to expansion of glaciers and formation of ice shelves (e.g., Ellesmere Island, Jones *et al.*, 1999; Mann *et al.*, 1998, 1999; Blake, 1989; Stewart and England, 1983). This cooling peaked 100 to 299 years ago (Bourgeois *et al.*, 2000), as recorded in ice cores from the Agassiz Ice Cap on Ellesmere Island (Williams *et al.*, 1995; Koerner and Fisher, 1990), and was followed by abrupt warming from 1900 to 2000.

Climate Change over the 20th Century

Although the current North American inter-glacial has been slightly cooler (2°) and wetter than the previous (~120,000 years BP), the last 100 years have been characterized by a rapid increase in temperatures, with global surface temperatures increasing by 0.3 to 0.6°C (IPCC, 1996), and Arctic temperatures rising by 3°C (Figure 1, Table 2). Climate changes over the 20th century have been attributed to high atmospheric concentrations of greenhouse gases, decreased solar irradiance, reduced volcanic activity, and changes in circulation (North Atlantic Oscillation, Arctic Ocean, Metcherskaya *et al.*, 2001) associated with (potentially anthropogenically-induced) changes in sea surface temperatures (McBean *et al.*, 2005).

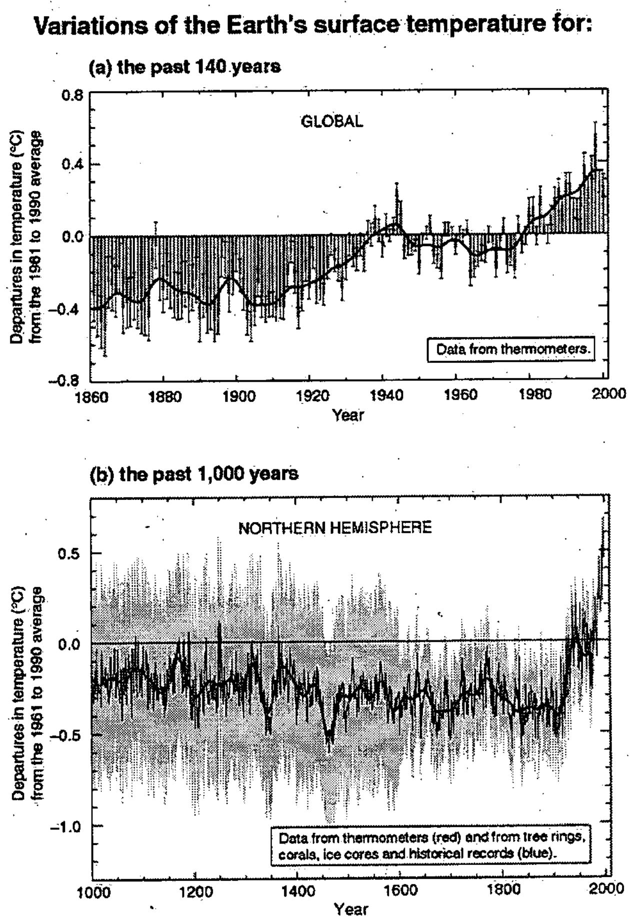


Figure 1. Surface temperatures in the Northern Hemisphere from 1000 yBP to 2000, and globally from 1860 to 2000 (IPCC, 2001, Figure 1, p.3). Temperatures were naturally variable over the last 1000 years, rising at an unprecedented rate and magnitude in the 20th century.

Table 2. Climate change over the 20th century (compiled from e.g., McBean *et al.*, 2005).

Parameter - Timeline	Details	Source
Surface air temperature		
1900-1945	Increase	
1946-1965	Decrease	
1966-2001	Increase	
1998, 2001, 2002	Three warmest years of instrumental record; similar to peak Arctic warming of 1930s	http://www.ncdc.noaa.gov/
	Greatest warming in winter and spring Trend toward warming over northwestern North America; cooling over eastern Canada > 1°C (>2°C winter) rise in Circumarctic temperature 1900-2000	Chapman and Walsh (1993)
Precipitation		
1900-1945	Increase over cold season and over summer in Canada	
1966-2001	Increase (up to 20% annually)	Groisman and Easterling (1994), Mekis and Hogg (1999)
	Shift towards more intense, more frequent	Stone *et al.* (2000)
Days with thaw (Temperature near -2°C; with snow cover)		Brown (2000)
Second half 20th century	Significantly increasing (1.5 to 2 day/50 year) for winter and autumn; 20% (winter) to 40%(autumn) increase in thaw frequency (North America and Russia)	
Heating degree-days (Positively correlated with energy consumption)		Guttman and Lehman (1992)
Second half 20th century	6% decrease over Circumarctic	
Frost-free period		
Second half 20th century	7% increase over most of Arctic 8% (or 9 days) in Eastern Canada Increase in length of frost-free period and degree-days Annual 'severity' of cold season decreased notably circumarctically, except in eastern Canada	Easterling (2002), Bonsal *et al.* (2001), Steurer and Crandell (1995)
Precipitation -Evaporation		
Early 1970s – early 1990s	Mean 16-17 cm (no trend)	Walsh *et al* (1994), Serreze *et al.* (1995)
Sea ice reduction		
1918-1938	2.9% ± 0.4% /10 years	Zakharov (2002), Cavalieri *et al.* (1997)
Snow Cover		
1972-current	10% decrease in area for N. Hemisphere	Groisman *et al.* (1994a)
1946-1995	Snow depth decline in winter; duration decline in summer	Brown and Braaten (1998)
1917-1997	1980s to early 1990s - rapid decline in areal extent Overall increase (all of North America)	Brown (2000)
Summary		
Air T	Increase (*winter and spring)	(Serreze *et al.*, 2000)
Spring snow extent	Decrease	
Spring SWE	Decrease	
Annual P	Increase (1900-1960) Increase (last 40 y N55° Canada)	

Paleoclimate records indicate that subsequent to the Little Ice Age, from around 1840 to the mid 1900s, the Arctic as a whole warmed at a rate and magnitude unprecedented over the previous 400 years (Overpeck *et al.*, 1997). Some areas of the Arctic warmed by 5°C over the 20th century alone (e.g., greatest warming in Alaska, central Canada and Siberia), while other regions cooled (e.g., Eastern Canada, North Atlantic, Greenland, Serreze *et al.*, 2000; Borzenkova, 1999a, b; Jones *et al.*, 1999). Generally, the greatest increases in air temperatures occurred in spring and winter, resulting in decreased snow cover, increased thaw (CCME, 2003), and a 20% increase in growing degree days (Table 2; Weller and Lange, 1999; Chapman and Walsh, 1993). Air temperatures in Canada's northeast, on the other hand, increased most greatly in summer and declined in winter over the past 50 years (CCME, 2003). Records from the Agassiz Ice Cap indicate that over this period of warming summer temperatures reached their highest in 1000 years (Koerner and Fisher, 1990).

Sediment records from lakes of arctic Canada reflect the recent warming associated with onset and progression of the industrial period. Ecological archives indicate that the climate of Nunavut was stable for many millennia and, like much of Canada, experienced a drastic shift in climate in the 1800s. Diatom and chrysophyte records from Char, Meretta, Sawtooth and Kekerturnak lakes of Canada's Arctic Archipelago changed remarkably over the 18th and 19th centuries (Wolfe and Perren, 2001; Douglas *et al.*, 1994), and more recently from 1988-1997 (Michelutti *et al.*, 2002). These changes were consistent with a longer ice-free, hence extended growing season. Lake diatom communities of western arctic Canada also shifted considerably over the past 150 years (Rühland *et al.*, 2003). Shifts in paleoclimatic proxies from Canada's eastern Arctic, on the other hand, have been much less pronounced, indicating continued climate stability in this region (e.g., Nettilling Lake, Baffin Island, Jacobs *et al.*, 1997; Saglek Lake, Labrador, Paterson *et al.*, 2003; northern Québec, Laing *et al.*, 2002).

Precipitation has increased in most areas of the Arctic since the 1950s (Groisman and Easterling, 1994; Groisman *et al.*, 1991). Increases in precipitation have occurred in all seasons over much of northern Canada (Table 2; IPCC, 2001a, 1996; Serreze *et al.*, 2000;

Zhang *et al.*, 2000; Groisman and Easterling, 1994), reaching up to 35% in some areas of the Arctic. Precipitation increases in arctic Canada have been most pronounced in autumn, and greatest from the Keewatin District to the central Arctic Archipelago, Nunavut (Environment Canada, 1995). Though precipitation increases at high latitudes over the 20th century were attributed to increased snowfall (CCME, 2003), snow cover and depth in the Northern Hemisphere declined substantially from the mid to late 1900s in association with reduced spring and summer snow covers and rising spring temperatures (Serreze *et al.*, 2000).

V. RECENT AND ONGOING CHANGES IN THE LANDSCAPE AND FRESHWATER RESOURCES OF THE ARCTIC

Rapid rise in annual temperatures, strong winter warming over the 20th century and even more so over the past 50 years, and increased precipitation have already resulted in many changes in the Arctic. These changes, though regionally variable (e.g., substantial warming in Central Russia, Alaska and Western Canada; cooling in Greenland; no clear trends in Nordic countries and northwest Russia, ACIA, 2005), generally include reduced sea ice extent and duration, retreat of glaciers and reduced snow cover, changes to freshwater ice and flow regimes, and melt of permafrost. For a comprehensive examination of recent changes to arctic landscapes and freshwaters in response to climate change, please see the ACIA (2005).

Sea Ice

Sea ice extent and thickness have declined in response to 20th century warming. Air temperatures over sea ice rose by 0.9°C between 1987-1997 (Aleksandrov and Maistrova, 1998), resulting in warming of marine waters (Carmack, 2000; Alekseev *et al.*, 1997; Kotlyakov, 1997; Carmack *et al.*, 1995) and contributing to a near 3% decrease in sea ice extent per decade (Johannessen *et al.*, 1999; Parkinson *et al.*, 1999; Serreze *et al.*, 1998; Cavalieri *et al.*, 1997), with accelerated loss in the 1990s (Maslanik *et al.*, 1996). Sea ice extent has declined by approximately 20% over the past 30 years in the North Atlantic (Johannessen *et al.*, 1999), and by near 5% in the Canadian Arctic (Anisimov *et al.*, 2001). Decreasing sea-ice extent and duration has been observed in Sachs Harbour, NWT, where, according to traditional observations, the 1990s were characterized by unprecedented changes in sea ice including: reduced quantity and thickness, and increased distance from shore, of multi-year ice; dramatic changes in the timing of spring break-up and autumn freeze-up with warmer winter and fall temperatures; more open water with more pronounced wave action in response to stronger winds and more ice movement; thinner and less extensive annual ice (Nichols *et al.*, 2004). The open-water season in Hudson Bay in the late 20th century was approximately 1 week longer than it was 30 years prior (CCME, 2003). In the eastern Canadian Arctic, where cooling was noted over the 1990s, sea ice extent increased (Baffin Bay, Labrador Sea, Serreze *et al.*, 2000; Chapman and Walsh, 1993), substantiated by increased sea salt concentrations in Penny Ice Cap (Grumet *et al.*, 2001).

Sea ice thickness has also declined over late 1900s, though the data are highly variable. The thickness of sea ice was determined to have decreased by up to 42% (3.1 to 1.8 m on average) along a transect from the Chukchi to Beaufort seas (Wadhams and Davis, 2000; Rothrock *et al.*, 1999). Other studies found much lower reductions in sea ice thickness over the last half of the 20th century (e.g., little change from 1970s to 1990s through Fram Strait, Vinje *et al.*, 1998; 5 to 7% per decade, Nagurnyi, 1995). Changes to sea-ice have had serious implications for coastal communities in the Arctic. For example, reduction in sea ice thickness and lengthening of the ice-free season in Alaska, together with permafrost melt, have resulted in increased shoreline erosion and damage to infrastructure (Berner *et al.*, 2005). As well, shoreline erosion has destroyed portions of Sarichef Island on the Chukchi Sea, limiting access of local communities to water and wastewater systems.

Ice Masses and Snow

Glaciers and ice caps of the Arctic have also responded to changes in air temperatures and precipitation. Though some glaciers have gained mass over the past few decades with moderate increases in precipitation, ice masses of the Arctic have generally experienced a net loss of mass with rising temperatures (Walsh *et al.*, 2005). Glaciers of the eastern Canadian Arctic lost mass from the early 1960s to the 1980s (Koerner, 1996), with further loss over recent decades. Melville Island South Ice Cap and White Glacier of Axel Heiberg Island declined in size over recent years, whereas Thompson Glacier grew (CCME, 2003; Koerner and Lungaard, 1995). The rate of retreat of the Barnes Ice Cap on Baffin Island, on the other hand, did not change significantly over the latter part of the 20th century, though mass loss was evident (Jacobs *et al.*, 1993) and the margin of the nearby Lewis Glacier retreated by approximately 680 m from the 1960s to the 1980s (Bell and Jacobs, 1997). Alaskan glaciers have lost thickness at a rate of 0.52 my^{-1} (Arendt *et al.*, 2002), and small arctic glaciers and perennial snow patches located at low altitudes have in some cases disappeared altogether (Anisimov *et al.*, 2001). Greenland's ice sheet has thinned significantly, losing mass on its southern and eastern margins (Krabill *et al.* 1999, 2000).

Though precipitation has increased over recent decades, dramatic increases in temperatures have resulted not only in decline in the thickness and extent of some perennial snow and ice, but also in the thickness and extent of annual snow cover. The extent of snow cover in the Northern Hemisphere has declined by 10% (near 61 260 km^2) since 1972, primarily due to enhanced spring and summer melt with rising spring air temperatures (Brown, 2000; Serreze *et al.* 2000). Though snowfall has increased substantially across northern Nunavut over recent decades (CCME, 2003), snow cover extent has declined with early spring melt in the west, and in locations such as Hall Beach, though other areas of Nunavut have noted a delay in spring melt (Maxwell, 1997; Foster, 1989). Paleolimnological records from proglacial Bear Lake on Devon Island (Lamoureux and Gilbert, 2004) indicate that underflows (related to autumn snowpack) over the 20th century were the most frequent and of the greatest magnitude over the past

750 years (Lewis *et al.*, 2002), indicating substantial snow accumulations over the last century. Sedimentation rates, however, declined in association with cooling and less pronounced melt over recent decades (e.g., Penny Ice Cap, Grumet *et al.*, 2001; Devon Island Ice Cap, Paterson *et al.*, 1977). Snow depth has also declined in Canada over the latter half of the 20th century, and frequency of rainfall and thunderstorms has increased (Nichols *et al.*, 2004).

Freshwater Discharges and Ice

River discharges have also changed over recent decades. A fifty year record indicated that from the late 1970s to the early 21st century winter runoff generally increased (Georgievsky *et al.*, 2002). This rise in quantity of runoff over the cold season was most pronounced in Russia and Siberia, increasing by 60% in some rivers. Annual discharge of large Eurasian rivers increased by 7% between 1936 and 1999 (Peterson *et al.*, 2002). Discharges of North American rivers have also increased, rising by 3 to 7% between 1921 and 1999, with exception of discharge into Hudson Bay, which decreased by 6%. Increases in river discharge over the last 30 years have occurred in association with rising air temperatures and increasing precipitation (e.g., Siberia, Alaska, New *et al.*, 2000). For example, runoff in the Back River, Nunavut, increased by 5.8 mmy^{-1} from 1966-1996, in conjunction with a 1.7°C rise in air temperature and a 5.5 mmy^{-1} increase in precipitation.

There is also evidence of recent changes in lake and river ice. Assessment of a 150 year period (1846-1995, Magnuson *et al.*, 2000) indicated that globally, on average, freeze-up has occurred 5.7 days late per century, and break-up 6.3 days early per century (i.e., duration of lake and river ice has declined by 12.3 d/100 years), associated with a 1.2°C increase in air temperature and substantive changes in spring and fall temperatures. Extension of the open-water season in lakes over recent years corresponds well with diatom records from Char Lake, Cornwallis Island, which show that from 1988 to 1997 (the warmest decade on record, with high snowfall) there was a pronounced shift in diatom assemblage toward increasing diversity and greater abundance of planktonic species, likely associated with a longer growing season (Michelutti *et al.*, 2003). Diatom records from Sawtooth Lake, Fosheim Peninsula on Ellesmere Island, also indicate a similar, dramatic change in diatom communities over this period (Perren *et al.*, 2003). Similar changes have been evident in other areas of the Arctic as well (Korhola *et al.*, 2002; Sorvari *et al.*, 2002).

Arctic rivers are generally freezing later and breaking up earlier (e.g., southern Finland, northern Sweden, Latvia, Kuusisto and Elo, 2000; Zachrisson, 1989). Rivers of northwestern North America are generally breaking up earlier (by 5 days per 100 years), and the interannual variability of their dates of break-up is increasing (e.g., Tanana River 1917-2000, Sagarin and Micheli, 2001; Yukon River 1896-1998, Jasek, 1998). Zhang *et al.* (2001), in reviewing trends in Canadian streamflow, found a similar advance in break-up of rivers in Canada's western Arctic. River freeze-up, however, generally occurred earlier, resulting in a net reduction in the length of the open-water season. Rivers of arctic and sub-arctic Russia and Western Siberia, on the other hand, have typically experienced a delay in freeze-up and a 7-10 day advance in break-up over the last century, increasing the length of the open-water season (Soldatova, 1993). The opposite is true for large rivers of Central and Eastern Siberia, which have experienced a slight advance in freeze-up and delay in break-up (Soldatova, 1993; Ginzburg *et al.*, 1992).

Permafrost

Changes to permafrost in the Arctic have also been evident, typically occurring in association with decadal variations in air temperatures, and varying highly regionally and temporally over the 20th century (Walsh *et al.*, 2005). Generally, circumarctic permafrost temperatures have risen over the past few decades. For example, surface temperatures in Alaska increased by 2 to 4°C in the 1980s (Lachenbruch and Marshall, 1986), and by 3° over subsequent years (Nelson, 2003). Permafrost temperatures of northwestern Canada increased slightly less so, by 2°C (Nelson, 2003). Changes to permafrost temperatures in northern Canada over the last two decades ranged from a decline of near 0.1°Cy^{-1} in

Northern Québec (Allard *et al.*, 1995) to an increase of approximately 0.15°Cy^{-1} in Alert, Nunavut (Geological Survey of Canada in Walsh *et al.*, 2005). Although air temperatures of eastern Canada (e.g., Ungava) have declined slightly, temperature of shallow permafrost has increased by almost 2° since the mid-1990s (Allard *et al.*, 1995; Brown *et al.*, 2000). Permafrost warming has been less pronounced in northwest Siberia (Pavlov and Moskalenko, 2002).

Variation in permafrost temperatures have been observed to depend on local soil and vegetation. Pavlov (1997) documented changes to permafrost temperatures over a 20 year period of warming (1979-1995, 2-2.5°C increase), in a variety of soils with variable ground cover (Table 3). The greatest increases in permafrost soil temperatures over the 20 year period occurred in the top 3 m of each soil type, and were most pronounced in landscapes with woody vegetation and nonvascular vegetation cover. Peatland and lightly vegetated tundra soils did not respond as rapidly to increasing air temperatures over the 20 years as did more heavily vegetated loam soils.

Table 3. Change in permafrost temperature with depth in response to warming over various ground covers and soil types at the Marre-Sale Station in Russia (adapted from Instanes *et al.*, 2005).

Ground surface and subsurface soil type	Depth (m)	Mean soil temperature (°C)	Soil temperatures increase 1979-1995 (°C)
Slope with willow and green moss cover; sand, loam	3	-5.4	2.2
	6	-5.3	1.2
	10	-5.2	0.8
Horizontal, hilly peatland with grass, shrub, moss and lichen cover; peat, ice and sand	3	-5.6	1.1
	6	-5.6	1.0
	10	-5.6	0.7
Polygonal tundra with moss, lichen, grass and shrub cover; sand	3	-6.5	1.3
	6	-6.5	1.1
	10	-6.4	0.6

Instanes and Mjureke (2002) analyzed freeze and thaw indices for various areas within the Arctic. The indices were derived from active layer thickness and permafrost temperature, providing a regional examination of unusually warm summers and winters over the last two decades (Tables 4 and 5). Records from Coral Harbour, Nunavut, indicate that 29% of summers from 1981-2000 were unusually warm (high thaw index), whereas 60% and 46% of summers were unusually warm at the NWT stations of Coppermine and Fort Smith, respectively. Trend analysis revealed a tendency toward "possible or weak cooling" at the Nunavut station, and a tendency toward "warming" at the NWT stations. Observed occurrences of unusually warm summers in Greenland were notably 0%, with a "cooling" trend. Unusually warm winters (low freeze index) accounted for 37% of the winters from 1981-2000 at Coral Harbour with no definitive trend toward warming or cooling (Coppermine 49%, warming; Fort Smith 55%, warming; Greenland 5 and 7%, cooling). Valdez, Alaska, had the highest % unusually warm winters (80) and summers (64). This analysis is consistent with observations of substantial warming in western North America, which contributed to loss of permafrost in central and western Canada and Alaska (Jorgenson *et al.*, 2001; Osterkamp *et al.*, 2000; Weller and Lange, 1999; Weller, 1998; Halsey *et al.*, 1995).

Table 4. Observed and expected occurrence of unusually warm summers at various stations in the Arctic, 1981-2000* (adapted from Instanes *et al.*, 2005).

Station	**Location**	**Observed 1981-2000 (%)**	**Expected (%)**	**Deviation (trend)**
Akureyri	Iceland	22	17	(+)
Ammassalik	Greenland	0	19	-
Anadyr	Russia	15	20	(-)
Barrow	Alaska	39	24	+
Bethel	Alaska	37	25	+
Coppermine	Canada	60	32	+
Coral Harbour	Canada	29	35	(-)
Fairbanks	Alaska	52	21	+
Fort Smith	Canada	46	24	+
Naryan Mar	Russia	33	26	(+)
Nome	Alaska	50	20	+
Nuuk	Greenland	0	14	-
Salekhard	Russia	35	17	+
Sodankylä	Finland	4	21	-
Svalbard airport	Svalbard	47	25	+
Turukhansk	Russia	20	17	**0**
Valdez	Alaska	64	24	+
Vardø	Norway	9	13	**0**
Verkhoyansk	Russia	40	17	+
Vilyuysk	Russia	34	19	+
Yakutsk	Russia	22	16	(+)

* + = warming, (+) = weak or possible warming, **0** = no trend, (-) =possible or weak cooling, - = cooling

Table 5. Observed and expected occurrence of unusually warm winters at various stations in the Arctic, 1981-2000* (adapted from Instanes *et al.*, 2005).

Station	Location	Observed 1981-2000 (%)	Expected (%)	Deviation (trend)
Akureyri	Iceland	19	17	0
Ammassalik	Greenland	7	19	-
Anadyr	Russia	18	19	0
Barrow	Alaska	46	24	+
Bethel	Alaska	31	25	(+)
Coppermine	Canada	49	32	(+)
Coral Harbour	Canada	37	35	0
Fairbanks	Alaska	28	20	(+)
Fort Smith	Canada	55	24	+
Naryan Mar	Russia	18	26	(-)
Nome	Alaska	33	20	+
Nuuk	Greenland	5	14	-
Salekhard	Russia	29	16	+
Sodankylä	Finland	15	21	(-)
Svalbard airport	Svalbard	26	25	0
Turukhansk	Russia	33	17	+
Valdez	Alaska	80	24	+
Vardø	Norway	21	13	(+)
Verkhoyansk	Russia	51	17	+
Vilyuysk	Russia	41	19	+
Yakutsk	Russia	50	16	+

* + = warming, (+) = weak or possible warming, **0** = no trend, (-) =possible or weak cooling, - = cooling

The increasingly frequent, unusually warm summers and winters in arctic Canada have occurred in correspondence with altered water balances in permafrost landscapes. For example, 1998 was a very warm year in Resolute, Cornwallis Island, with a longer warm season (e.g., spring melt 1 month early), hence increased duration and extent (+0.01 to 0.02 m) of ground thaw, and enhanced evaporation (Young and Woo, 2003). These events were accompanied by a decline in precipitation. These changes resulted in a nearly irreversible reduction in the size of a semi-permanent snowbank, increased ground ice meltwater flows, and ground surface subsidence (e.g., 0.13±0.7m). Though wetland evaporation increased and water tables lowered with high temperatures and low precipitation, and ground thaw extended to deeper depths, water tables and runoff at some sites were sustained by the melt of semi-permanent snowbanks and ground ice.

Nunavut Case Study

Recent observations of climate change in Nunavut were compiled through a research project that began in 1995 based upon the communities of Iqaluit, Igloolik, Baker Lake (Qamani'tuaq) and Clyde River (Kangiqtugaapik) (Huntington *et al.*, 2005b). These communities have observed many changes in recent years. These changes include increased variability and decreased predictability of weather, greater extremes, and added travel hazards including deteriorating ice and snow conditions. These changes in weather have been noted since the early 1990s and are consistent with observations from other areas of arctic Canada and Alaska (Reidlinger *et al.*, 2001; Whiting, 2002). Other observations included a gradual decline in water levels of lakes and rivers around the Baker Lake community since the 1960s, with dramatic reduction in water levels in the 1990s, progressing to severe in 1998-2002. These lower water levels caused some shallow lakes and rivers to dry, reduced water quality, blocked travel routes, and had a detrimental effect on fish abundance, health and quality (e.g., L. Arngaa'naaq, Baker Lake, 2001; N. Attungala, Baker Lake, 2001).

VI. FUTURE CLIMATE CHANGE IN NUNAVUT

Modeling Future Climate

Future climate may be simulated through the use of atmospheric or oceanic (or coupled atmosphere-ocean) general circulation models (GCMs) that incorporate various future emission scenarios. Atmospheric-Oceanic General Circulation Models (AOGCMs) mathematically simulate the physical dynamics of the atmosphere and the ocean on a three-dimensional grid over the globe and over various time steps (Källén *et al.*, 2005). More sophisticated AOGCMs also model other components of the climate system, such as the land surface and cryosphere. Scenarios of future climate are generated by forcing these models with emission scenarios (i.e., plausible future greenhouse gas and aerosol levels). The performance of climate models may then be evaluated by comparing records of past and present climate with simulated reconstructions (Figure 2).

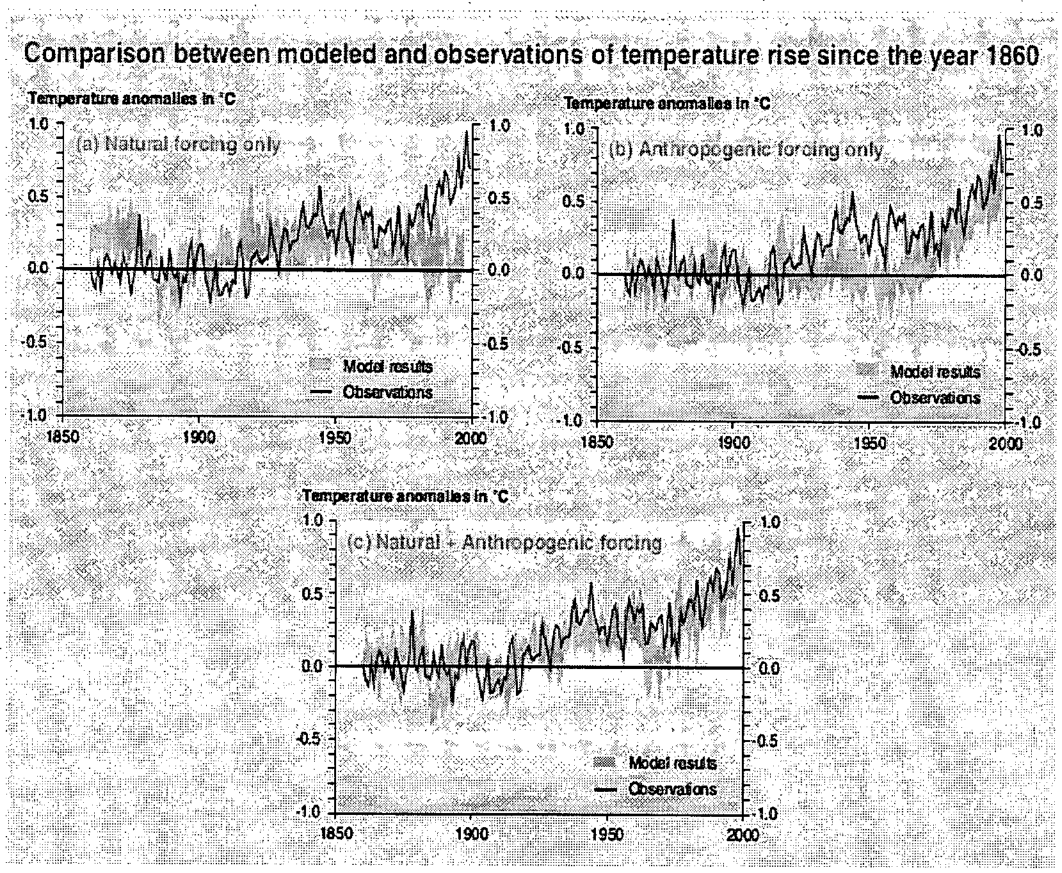


Figure 2. Modeled and observed temperatures between 1850 and 2000 (IPCC, 2001a, Figure SPM-2, p.7). The inclusion of natural and anthropogenic forcing produces a reasonably good simulation of observed temperatures over the past 150 years.

Huntington *et al.* (2005b) summarized approaches to choosing scenarios and models of climate change. Scenarios of future climate are based on estimation of future emissions of greenhouse gases and pollutants (IPCC-TGCIA, 1999). Various emission scenarios exist; those from the SRES (Special Report on Emission Scenarios, Nakienovi *et al.*, 2000) are considered equally valid pictures of future emissions of e.g., greenhouse gases. The B2 emission scenario (Figure 3), used in the ACIA (2005) and by the IPCC (2001), is based on intermediate levels of economic growth, continuously increasing population, moderate and diverse change in technology, and includes a diverse world with local differences.

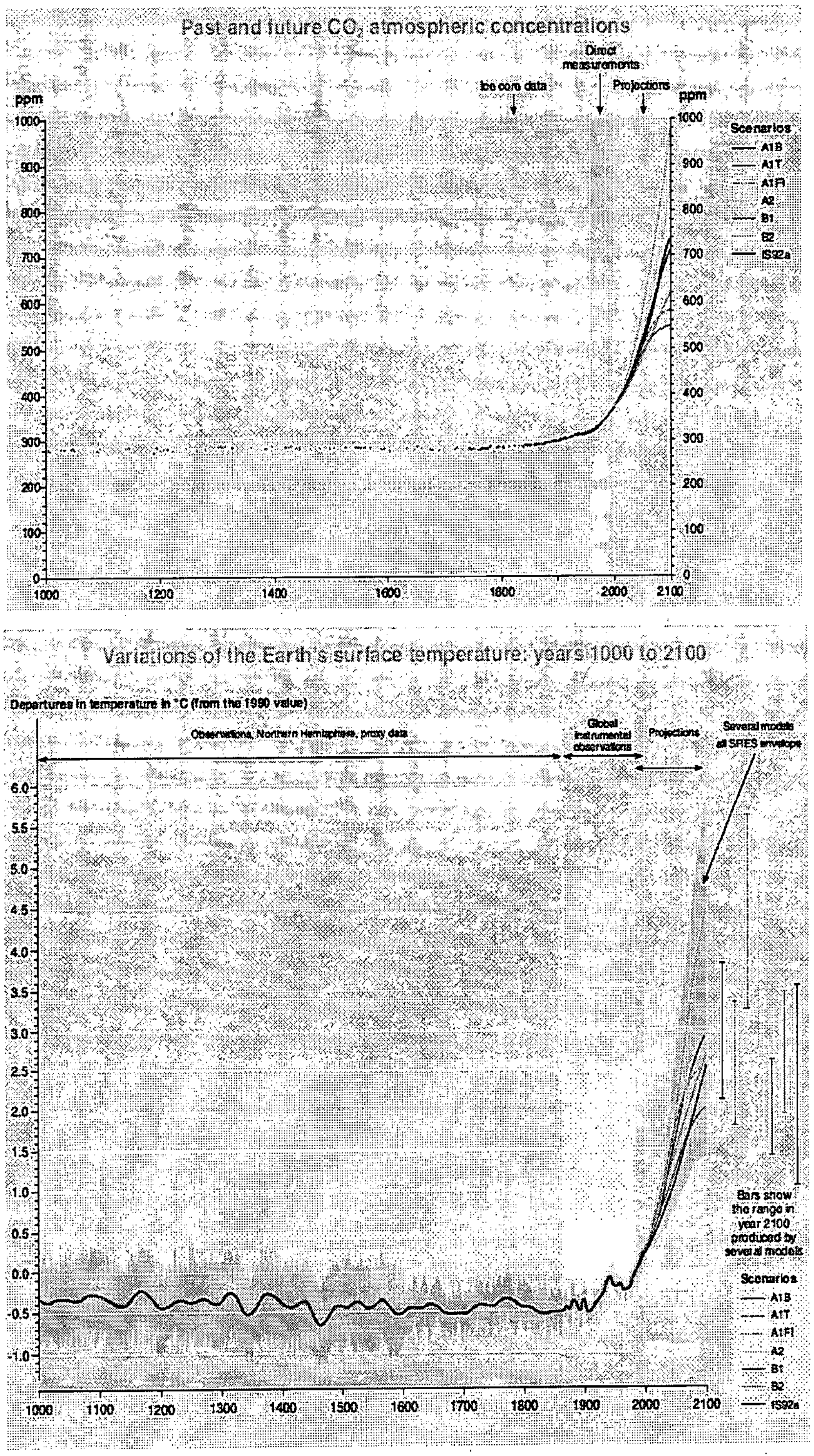


Figure 3. Historical and projected average global concentrations of atmospheric carbon dioxide, and average Northern Hemisphere surface temperature departure from 1990 mean. Projected future increases are based on six SRES emission scenarios, of which the B2 is intermediate (IPCC, 2001a, Figures SPM-10a,b, pp. 33-34).

This report on future climate change impacts on Nunavut relies heavily upon the climate change and impacts projections of the ACIA (2005). Multiple AOGCMs, forced by the B2 scenario, were used in the ACIA to provide a good representation of the potential range of future climates (IPCC, 2001) These five models were the CGCM2 (developed by the Canadian Centre for Climate Modelling and Analysis), CSM_1.4 (National Center for Atmospheric Research, USA), ECHAM4/OPYC3 (Max-Planck Institute for Meteorology), GDFL-R30_c (Geophysical Fluid Dynamics Laboratory, USA), and HadCM3 (Hadley Centre for Climate Prediction and Research, UK).

ACIA circumarctic and regional assessments of climate change impacts were based on the data generated by the five AOGCMs. This data included temperature, precipitation, evaporation, atmospheric pressure, cloudiness, shortwave radiation receipt, sea-ice, snow cover and length of the growing season. Projections were made to 2100, with 1981-2000 as the baseline climate. Projected changes in climate and impacts were described over three time slices (2020, 2050 and 2080 ± 10 years) and over four regions of the Arctic. Nunavut lies within Region IV, along with western Greenland. This spatial subdivision of the Circumarctic allowed for a regional analysis of climate change impacts in the Arctic, taking into account regional differences in projected climate change and impacts (e.g., local differences in the physical, biological and chemical structure and function of terrestrial and aquatic systems, maritime influences, communities, industry).

Future Climate Change in Nunavut and the Circumarctic

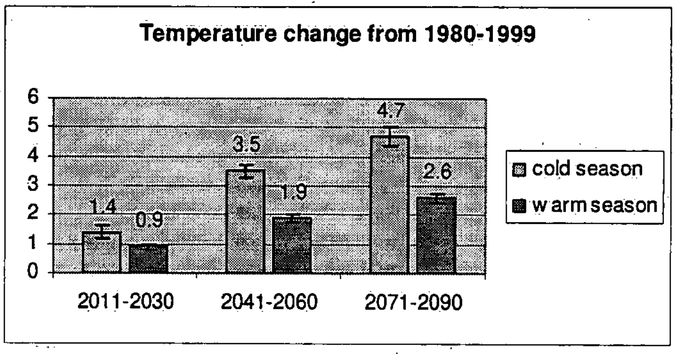
The five AOGCMs applied in the ACIA project a substantial rise in temperature and a moderate increase in precipitation in the Arctic over the next 100 years (Källén *et al.*, 2005). This is consistent with IPCC (2001a) projections of increasing temperature and precipitation associated with rising atmospheric carbon dioxide concentrations (Figure 3). This section examines projected changes in temperature and precipitation for the Circumarctic as found by the ACIA, and as tailored to Nunavut. Summaries of projected changes in temperature, precipitation and other climate variables are presented in Table 6 and Figure 4. The discussion that follows focuses on the seasonality and regionality of projected changes in temperature, precipitation, and other climate variables to 2071-2090.

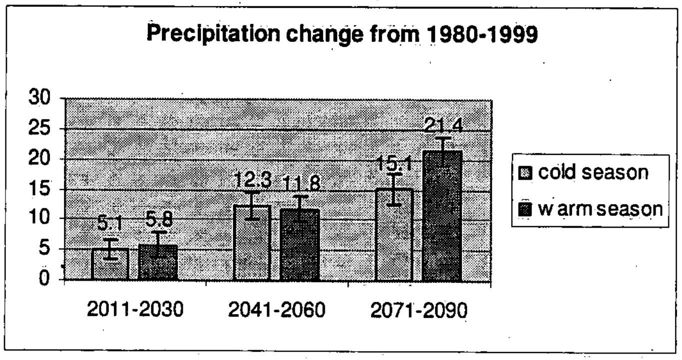
Table 6. ACIA-projected change in climate conditions of the Circumarctic (60-90°N) from 1981-2000 baseline to 2071-2090. Average, minimum and maximum projected changes are from the 5 ACIA models (CGC, CSM, ECH, GFD, HAD). Air temperatures are expected to increase, as is precipitation, and water availability (P-E) (from Källén *et al.*, 2005).

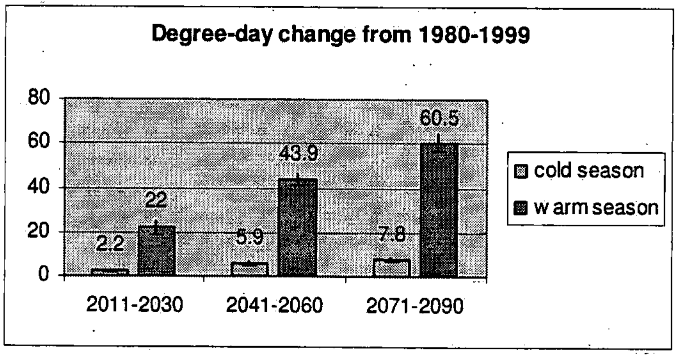
Primary Climate Variables and Derivatives	**Average Projected Annual Change**	**Minimum Projected Annual Change**	**Maximum Projected Annual Change**
Air temperature	+ 3.7°	+ 2.8° CSM	+ 4.6° ECH
Precipitation	+ 12.3%	+ 7.5% CGC	+ 18.1% ECH
Precipitation minus Evaporation (P-E)	+ 15.5%	+ 2.4% CGC	+ 30.2% ECH

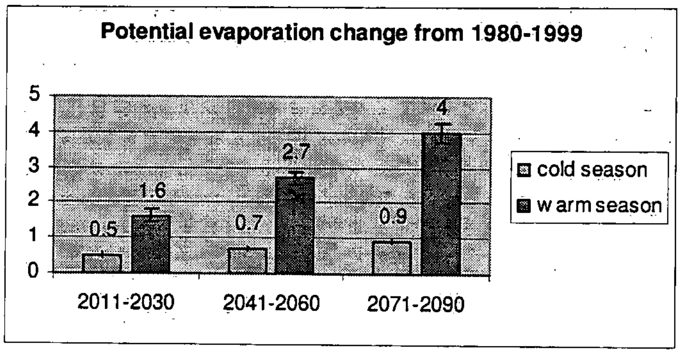
For the purposes of this report, ACIA GCM grid data, used in projecting future changes in arctic climate, were reexamined for the territory of Nunavut. Changes to temperature and precipitation, projected to 2020, 2050 and 2080, were recalculated for the landmass of Nunavut (excluding surrounding marine waters; Figure 4, Bill Chapman, unpublished data). Future changes in degree-day and potential evaporation were generated as well. The projected changes in climate were subdivided into warm and cold season values, averaged for June to September and October to May, respectively. These data, along with the projected changes in circumarctic temperature and precipitation (Figures 5 to 10), are useful in identifying seasonal and spatial trends in projected climate change for the territory of Nunavut.

Figure 4. Projected change in climate conditions of Nunavut from 1981-2000 baseline to 2011-2030, 2041-2060, 2071-2090. Data are cold (October to May – in blue) and warm (June to September – in purple) season averages for the landmass of Nunavut. Error bars are the standard deviation from the mean of the 5 ACIA models. The projected change in cold season temperatures (°C) over the 21st century is much greater than that of the warm season, while projected increases in precipitation (mm) are similar between seasons. The number of degree days (days with minimum temperature >0°C) also increases, as does potential evaporation (mm water that could be evaporated if available). (Bill Chapman, unpublished data)









Air Temperature

Overall, average annual air temperatures in the Arctic are projected to rise by 3.7°C by the late 21st century, increasing at almost twice the global rate of 1.9°C over the next 100 years (Källén *et al.*, 2005; global change of 3°C projected by IPCC, 2001a). Increases in arctic air temperatures over the past 50 years alone have been substantial, with distinct inter- and intra-regional, as well as seasonal, variability (Table 7). The greatest warming occurred in northwestern North America, where annual temperatures rose by 2-3°C, and by nearly twice that during the cold season. Areas in and adjacent to the northern North Atlantic cooled by 1-2°. Similar regional and seasonal variability in air temperatures is projected under future climate scenarios, with the recent trend toward rapidly increasing temperatures continuing through the 21st century (Table 7, Figure 3b).

The images presented in the discussion below (Figures 5-8) illustrate the projected change in Arctic air temperatures from 1980-1999 to 2070-2089, as determined by the ACIA (http://www.acia.uaf.edu/). Change in surface air temperature is the 5-model, monthly mean for winter, spring, summer and fall months. Units are in degrees Celsius, with + (yellow-red scale) indicating an increase in temperature, and – (green-blue scale) indicating a decline in temperature. The greatest increases in temperatures are expected over the cold season, spring, and autumn, with more pronounced warming over ocean and seas and coastal areas than inland. Statements regarding the regionality of projected changes within Nunavut are intended to provide a general indication of potential future climate conditions, and as such must be applied judiciously, particularly at local scales where interactions and feedbacks between climate system components will be highly variable, increasing uncertainty in the direction and magnitude of projected future climate change and impacts.

Table 7. Average change in air temperatures for each of the 4 ACIA regions annually and seasonally (winter) over the past 50 years and projected by the 5 ACIA models to the 2090s (data compiled from Källén *et al.*, 2005).

ACIA Region	**Change Over Past 50 years**		**Change from 1981-2000 baseline to 2090s**			
	Average annual air temperature	Average winter air temperature (inland)	Average annual air temperature (inland)	Average winter air temperature (inland)	Average annual air temperature (oceanic)	Average winter air temperature (oceanic)
I **E Greenland, Iceland, Fennoscandia, NW Russia and adjacent water**	+ 1° - 1° North Atlantic	+ 2-3°	+ 2-3°	+ 3-5° moreso over NW Russia and coastal areas	+ 6°	+ 6-10°
II **Siberia**	+ 1-3°	+ 3-5°	+ 3-5°	+ 3-7°	+ 10°	+ 10°
III **Chukotka, Alaska, W Canadian Arctic, adjacent water**	+ 2-3° Alaska and Yukon + 0.5° western region	+ 3-5° Alaska and Yukon + 1-2° Chukotka	+ 3-4°	+ 4-7°	+ 6°	+ 10°
IV **Central and E Canadian Arctic, W Greenland, adjacent water**	+ 1-2° - 1° Labrador Sea and adjacent areas of Canada and Greenland	+ 3-5° - 1-2° Labrador Sea and adjacent areas of Canada and Greenland	+ 3-5° Canadian archipelago	+ 4-7° Canada + 3-5° Greenland	+ 5-7°	+ 8-10° Hudson Bay, northern Labrador Sea, Arctic Ocean

+ indicates warming; - indicates cooling.

Winter air temperatures are projected to change substantially by 2071-2090 (Figure 5). Temperatures are expected to increase by less than 3°C in and adjacent to the North Atlantic, and by greater than 9°C over large water bodies such as the Arctic Ocean, Hudson Bay and the Labrador Sea (see Table 7 for projected changes in winter temperatures for the ACIA regions). The projected increase in temperatures over large water masses in the Arctic will extend to adjacent landmasses, where coastal temperatures in Russia and Canada are projected to increase by near 9°C, and inland temperatures by between 3-6°, for the most part. The extent and magnitude of warming inland dissipates through January and February as the severity of warming over the Arctic Ocean lessens. Temperature changes in Nunavut over this period will likely be greatest in December and January (near 6°C), particularly at higher latitudes, and will be less pronounced in February (3-6°C). Warming over Ellesmere Island is expected to be moderate, while more severe warming is projected for Southampton Island and southern Baffin Island (near 8-10°C) in association with projected warming over Ungava Bay and Labrador Sea. Overall, the average projected change in cold season (October - May) temperature over terrestrial Nunavut is + 4.7°C (Figure 4).

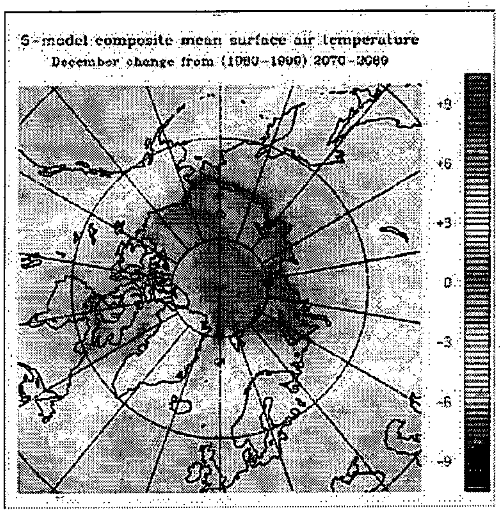
The substantive increase in winter air temperatures projected for December to February (Figure 5) changes to moderate warming through late winter and early spring (Figure 6). Surface air temperatures are expected to increase more substantially over marine and oceanic areas (6°) than over terrestrial locations (3-6°), with moderate warming inland and greater warming in coastal areas, particularly in Eurasia and the Canadian Arctic Archipelago (e.g., March). By May, projected increases in temperature are nearly equivalent over land and water masses, with little more than 3° warming of air temperatures by the late 21st century. Increases in temperatures in Nunavut are expected to be greater in late winter in the central Archipelago and southern Baffin Island (by up to 6°), and less pronounced and more uniformly distributed into early spring (<3°).

The magnitude of changes in projected air temperatures continues to decline into the warm season (Figure 7). Temperature changes projected for June through August are very different from those for winter and spring months (Figures 5 and 6). Average temperature changes are greater over landmasses rather than water bodies, and increases in air temperatures over oceanic and marine locations are expected to be very small. Slight cooling is projected for the northern Canadian Archipelago and the central Arctic Ocean (July). The greatest increase in warm season temperatures (near 3°) is expected for Siberia, central Greenland, and central Nunavut. Warm season temperatures over the northern islands of Nunavut are projected to change very little over the 21st century. Warming over Nunavut lessens and is more uniformly distributed over the territory into August, when oceanic warming increases relative to earlier months. Overall, the average projected change in warm season (June – September) temperature over terrestrial

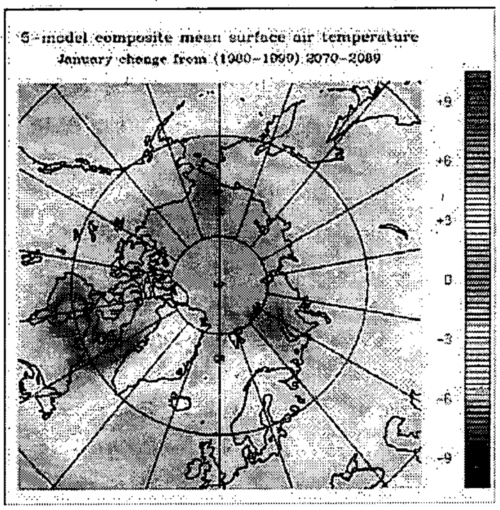
Nunavut is + 2.6°C (Figure 4), associated with a 60.5 day increase in degree days and a 4 mm increase in potential evaporation from baseline (1981-2000) to 2071-2090.

Projected changes in temperatures over oceanic and coastal locations progressively increase through the fall and early winter months (Figure 8). The months of October and November are projected to have the most severe increases in air temperatures over water masses by 2070-2089, and near the greatest seasonal warming of coastal and inland locations over the entire annual cycle of temperatures. Air temperatures are projected to increase by much greater than 9° over much of the Arctic Ocean by October, with increases in coastal air temperatures reaching near 9° over the northern Canadian Archipelago, with less substantial terrestrial warming over coastal Eurasia (>6°), and moderate warming of at least 2-3° over the rest of the Arctic. The severity and extent of warming over the Arctic Ocean is projected to peak in November, with air temperatures over the northern Canadian Archipelago increasing by greater than 9°, and more pronounced coastal and inland warming extending through Eurasia and the rest of northern North America. Warming is projected to be near the most severe (near 10°) of the entire terrestrial Arctic over the Queen Elizabeth Islands in October, extending inland in November. Inland warming is projected at just slightly greater than 3° over central Nunavut and Baffin Island in October. As such, a steep, latitudinal gradient in temperatures may be expected over the fall and early winter under projected future climate conditions.

a) December



b) January



c) February

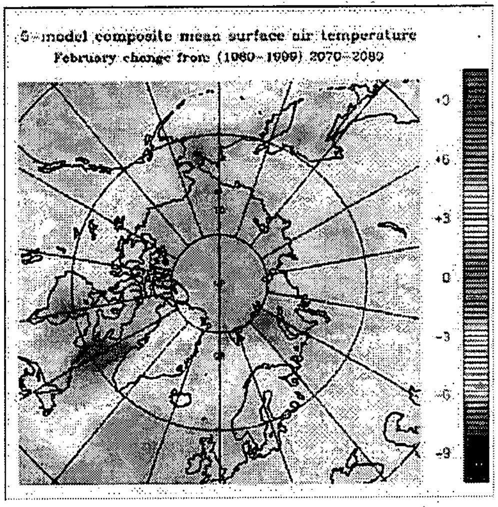
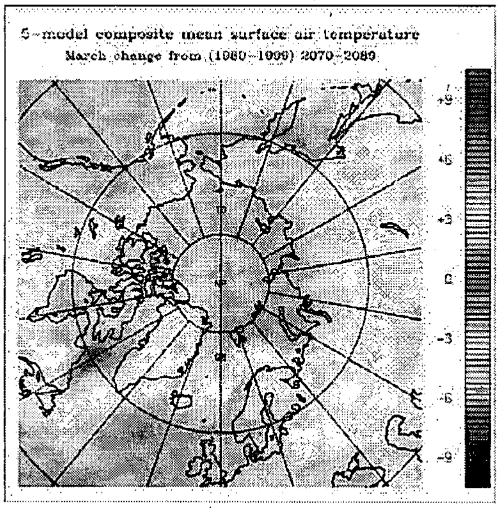
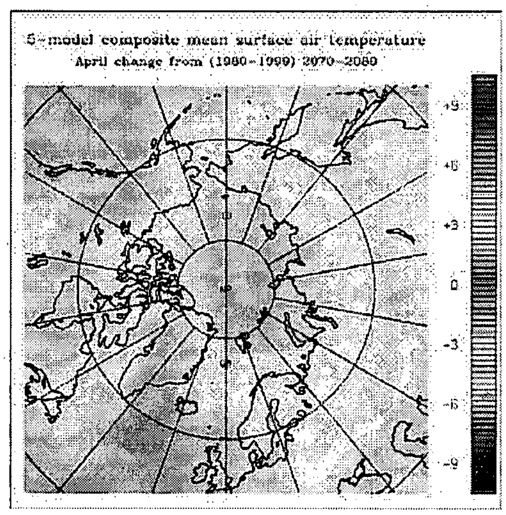


Figure 5. Change in a) December, b) January and c) February air temperatures to 2070-2089. 5-model mean of change from 1980-1999 (ACIA, unpublished).

a) March



b) April



c) May

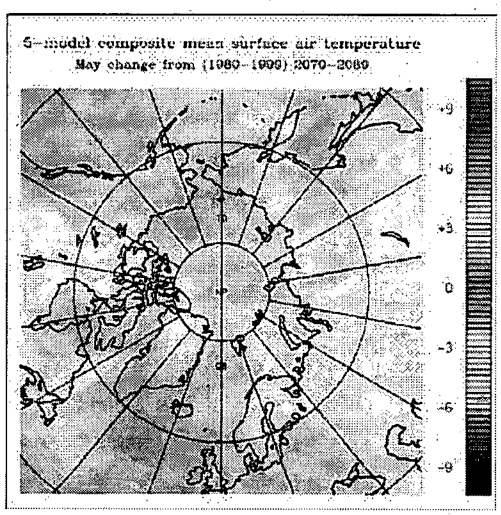
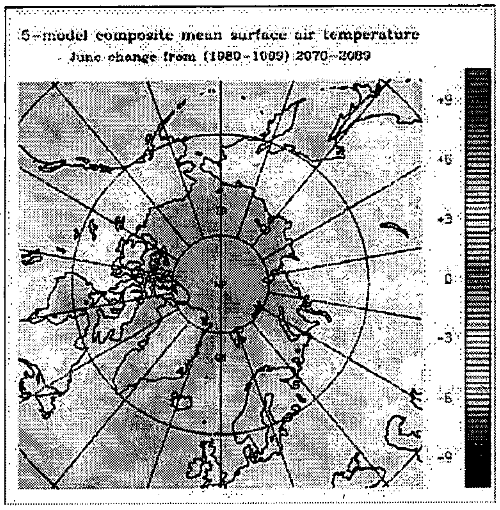
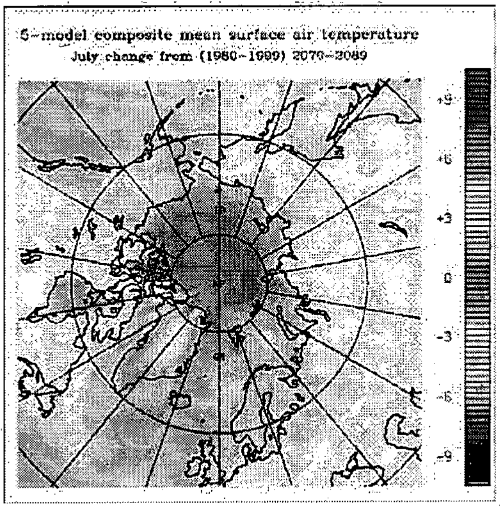


Figure 6. Change in a) March, b) April and c) May air temperatures to 2070-2089. 5-model mean of change from 1980-1999 (ACIA, unpublished).

a) June



b) July



c) August

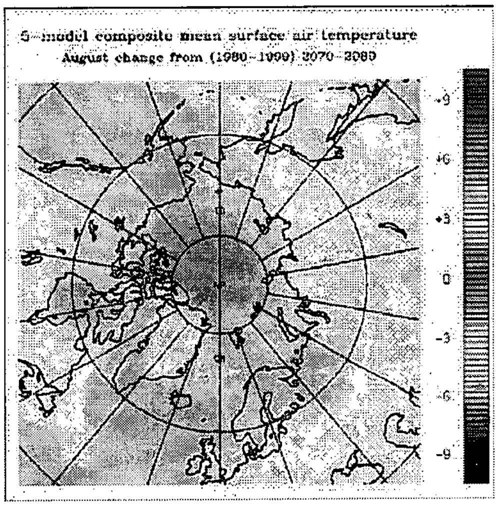


Figure 7. Change in a) June, b) July and c) August air temperatures to 2070-2089. 5-model mean of change from 1980-1999 (ACIA, unpublished).

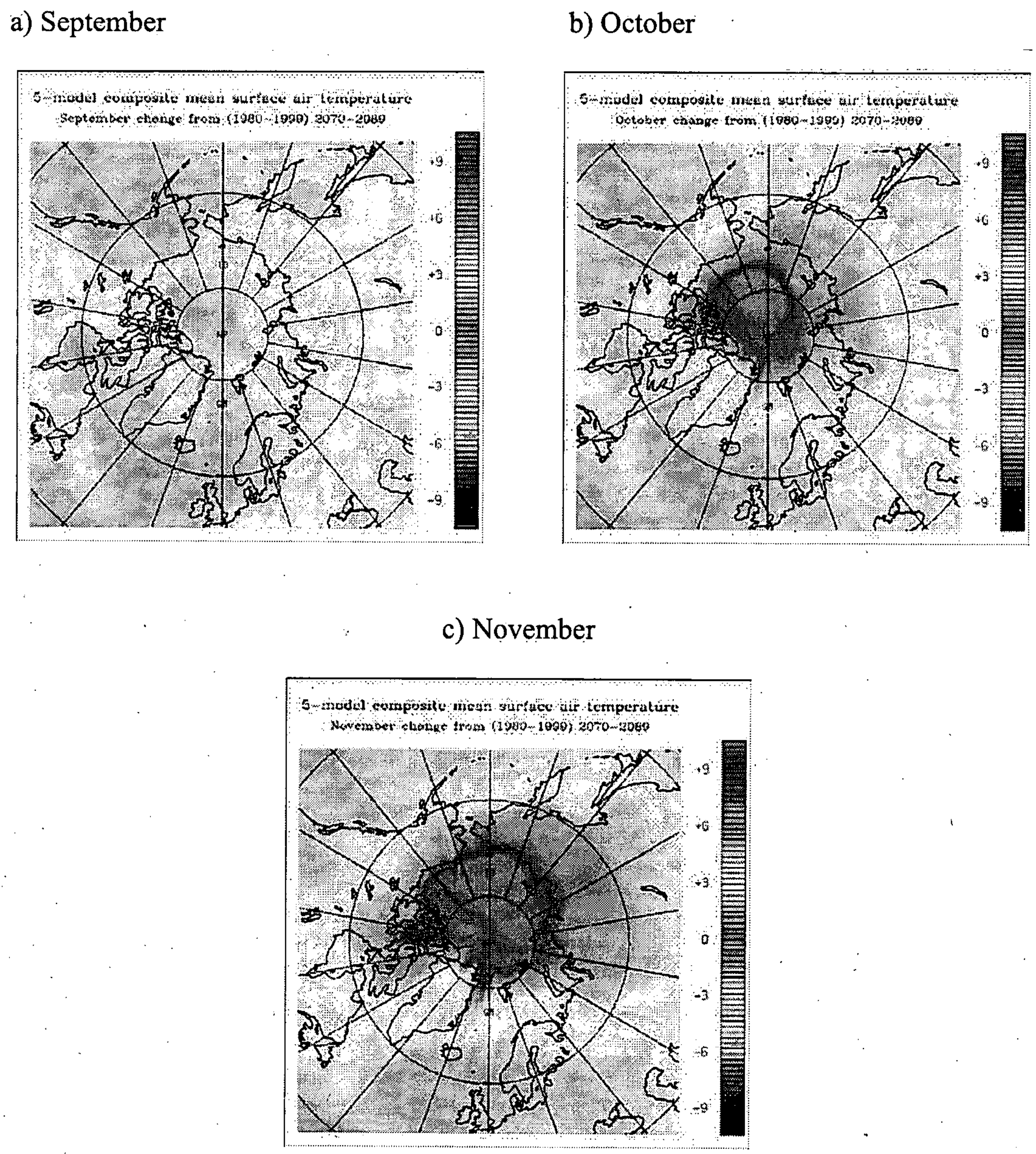


Figure 8. Change in a) September, b) October and c) November air temperatures to 2070-2089. 5-model mean of change from 1980-1999 (ACIA, unpublished).

Precipitation

ACIA AOGCM output indicate that globally, precipitation may be expected to increase on average by 2.5% by 2080, with an average increase of 12.3% for the Arctic (Table 6). This likely increase in precipitation at high latitudes is attributed in part to an expected enhancement of atmospheric circulation under a warmer temperature regime (e.g., Manabe and Wetherald, 1975), and to a decline in ice cover, hence increased evaporation and precipitation in Arctic locations (Walsh *et al.*, 2005; Källén *et al.*, 2005). Increases in precipitation under future climate scenarios are expected to be most pronounced in autumn and winter, and highly variable inter-annually. The magnitude and frequency of precipitation events are projected to increase, with potentially greater frequency of extreme events.

Changes to precipitation under future climate forcing will likely be highly variable regionally, as well as locally, under the influence of e.g., topographic effects, proximity to large water bodies. Mid-winter circumarctic precipitation (Figure 9a) is generally projected to increase over the entire Arctic (>6 mm/month), with a decline in precipitation over the Labrador Sea and the North Atlantic. Precipitation may increase by up to >12 mm/month over Hudson Bay, southern Alaska and the Bering Strait. A similar spatially heterogeneous scattering of precipitation changes is apparent in mid-summer as well (Figure 9b), with a trend toward declining precipitation (near -6 mm/month) over terrestrial Europe and Russia, as well as over the North Atlantic. Substantial increases in precipitation (>12 mm/month) are projected for Labrador, coastal areas along the Bering Strait, and isolated areas in Siberia and Russia. Precipitation changes in Nunavut to 2070-2089 are projected to be very slight for the most part in mid-winter, and slightly greater (near +4-6 mm/month) over much of Nunavut in mid-summer, particularly on the southeastern edge Ellesmere Island.

a) January b) July

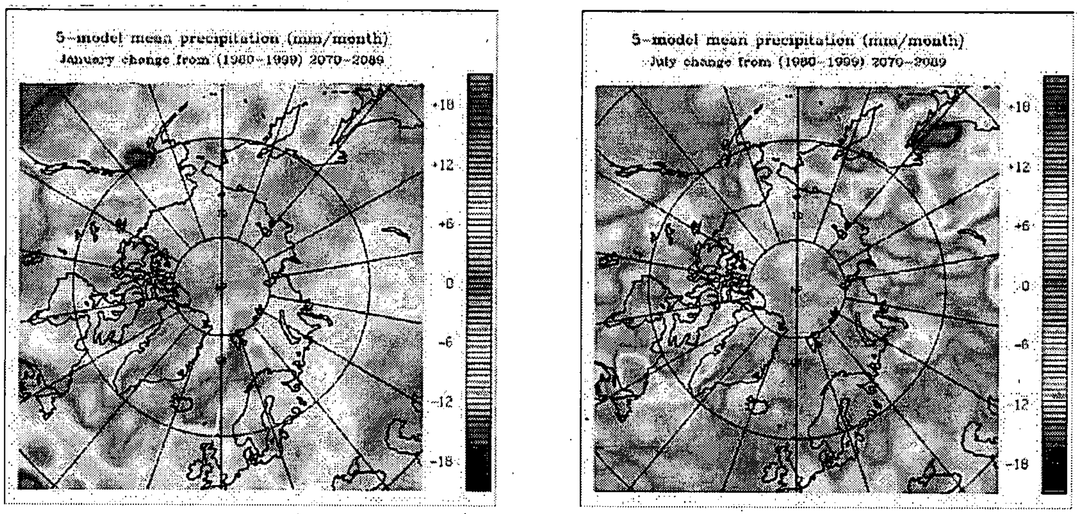


Figure 9. Change in a) January and b) July precipitation (mm/month) to 2070-2089. 5-model mean of change from 1980-1999 (ACIA, unpublished).

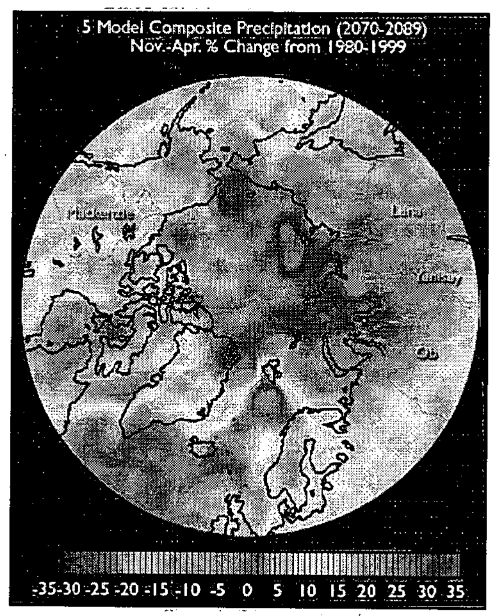


Figure 10. Percent change in cold season precipitation (November to April) to 2070-2089. 5-model mean of change from 1980-1999 (Wrona *et al*, 2005, Figure 8.10).

Composite % change in winter precipitation indicates that average winter precipitation in the Arctic is projected to increase for the most part by near 10 to 20% (Table 6, Figure 10). The greatest increases in precipitation are projected to occur over the Arctic Ocean, reaching near 35% near the Asian landmass, potentially resulting in >15% increase in precipitation for much of arctic Russia and Siberia. Increases in precipitation for arctic Europe, Greenland and North America are projected to be less pronounced, ranging from little change in arctic Europe and parts of Greenland, to increases of >20% in isolated areas of Alaska, Greenland, Quebec, and northernmost Canada. Precipitation change is not only nearly absent in the

North Atlantic, but negative to near 5%. The greatest change in terrestrial winter precipitation in Canada is projected for the northernmost portion of Nunavut (northern Queen Elizabeth Islands, >+25%) and the land adjacent to Foxe Basin (southwestern Baffin Island, Southampton Island, Melville Peninsula). Increases in precipitation are heterogeneously distributed over Nunavut to 2070-2089, but are generally less pronounced inland (10 to 15%) and at lower latitudes.

VII. PROJECTED IMPACTS OF FUTURE CLIMATE CHANGE ON THE LANDSCAPE AND WATER RESOURCES OF NUNAVUT

This section addresses the potential impacts of rising temperatures and increased precipitation on Nunavut over the 21st century. These impacts were assessed by the ACIA (2005) based on the above-mentioned GCM projections of temperature and precipitation, scientific research and knowledge of interactions between the components of the climate system and, where available, process-based models (e.g., soil models). The complex nature of interactions, variability of potential local responses, feedbacks, and the potential for cumulative and synergistic effects, as well as threshold responses, limit the certainty with which some of these projections may be made. See the ACIA (2005) for further discussion of this topic. The ACIA findings have been examined, and issues specific to Nunavut identified and discussed for the purposes of the current report. Where available, documentation specific to the territory of Nunavut has been provided. This section begins with an examination of climate change impacts on sea ice, glaciers and ice sheets, snow, permafrost, and river and lake ice as presented in Chapters 6, 8 and 9 of the ACIA. The discussion continues with potential impacts on water quality and infrastructure, as drawn from Chapters 6, 8, 15 and 16 of the ACIA.

Sea Ice

Background

Sea ice is a critical component of the climate system in the Arctic. Sea ice is not only a manifestation of the extremes in temperature at high latitudes and a sensitive indicator of climatic change, but is also a significant control of climate, affecting the surface energy balance of marine waters, the moisture budget of both marine and terrestrial areas, and the magnitude and intensity of coastal storms and erosion (Walsh *et al.*, 2005). Feedbacks between the ice surface and the climate system respond not only to changes in temperature, which can alter the thickness, extent and duration of sea ice, thereby affecting the thermal regime of marine ice and waters, but also to changes in precipitation. Snowfall may increases the albedo of the ice surface, positively feeding back to ice growth. Considerable snow accumulations may, however, alter the thermal balance of the ice through provision of an insulating layer, thereby limiting the thickness to which ice will develop. The extent and duration of sea ice, in turn, affects evaporative loss of water and generation of precipitation, as well as the occurrence of storms and storm surges, which are important to coastal land and water resources. This topic is discussed more thoroughly in Chapters 6 and 9 in the ACIA (2005).

Recent Trends

The areal coverage of arctic sea ice varies annually from near 5-6 million km^2 in September to near 14 million km^2 in March (Figure 11, Parkinson *et al.*, 1999). Sea ice extent has decreased by near 2.8 ± 0.3% per decade from the 1970s to the 1990s, corresponding with rising air and water temperatures (e.g., Carmack, 2000; Aleksandrov and Maistrova, 1998; Alekseev *et al.*, 1997). Decreases in sea ice extent have been greater in the summer than in the winter, consistent with indigenous observations in western arctic Canada and Alaska (Krupnik and Jolly, 2002), and have reached near 2% in Hudson Bay and near 2-3% in David Strait and the Labrador Sea (Chapman and Walsh, 1993). The coverage and thickness of multi-year ice have also changed, having decreased by near 9% per decade, with losses near 10 to 20 % within the Archipelago between the 1980s and 1990s (Comiso, 2002; Johanessen *et al.*, 1999). Ice thickness in the central Arctic Ocean has declined as well, by up to 40% or 1.3 m over the mid to late 1990s (e.g., Rothrock *et al.*, 1999; Vinje *et al.*, 1998; Nagurnyi, 1995).

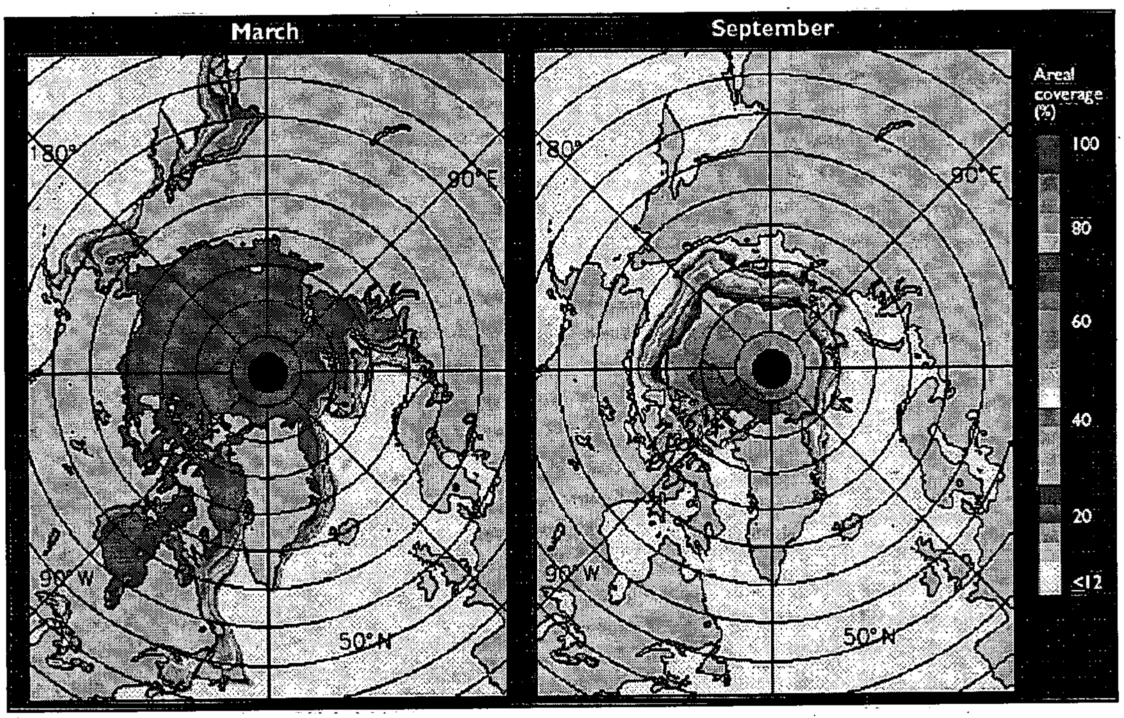


Figure 11. Mean concentration (percent areal coverage) of Northern Hemisphere sea ice in March and September, 1990-1999 (C. Parkinson, 2003 in Walsh *et al.*, 2005, Figure 6.3). At it's maximum areal extent (in March), sea ice dominates northern Nunavut, and has slightly less coverage in Hudson Bay, Ungava Bay and Davis Strait. At it's minimum areal extent (in September), sea ice is absent from much of Nunavut, with % coverage in the archipelago increasing with latitude, and channels of the Queen Elizabeth Islands, Viscount Melville Sound, and McClintock Channel typically not clearing completely over the warm season (Maxwell, 1997).

Future Changes

The ACIA projects a decline in sea ice extent through the 21st century (Walsh *et al*, 2005). The projected decline in mean annual extent of sea ice in the Northern Hemisphere from 2000 to 2100 varies from 12% (from 12.3 to 10.8 x 10^6 km^2, CSM) to

46% (from 12.3 to 6.6 x 10^6 km^2, CGC). The wide range in values of projected sea ice extent is a product of the highly variable nature of the data input to each model (e.g., initial ice thickness values vary between models). The most notable decline in ice extent is projected for the fall, with little change expected over the winter (Figure 12). The decline in fall (September) ice extent in Nunavut is projected to be minimal to 2010-2030 relative to 1990-1999 (Figure 11). By 2050, the coasts of Nunavut's mainland are expected to be largely ice-free in the fall. By 2080, fall sea ice is projected to be present only in the northernmost Queen Elizabeth Islands and along the northern coast of Ellesmere Island. Other studies have estimated a decline in future summer sea-ice extent of 60 to 80%, with coasts remaining ice-free nearly twice as long and ice retreating to distances nearly 500-800 km from shore (Vinnikov *et al.*, 1999; Gordon and O'Farrell, 1997). Though winter sea ice is projected to remain widely distributed over the Arctic, it is expected that there will be retreat of the ice edge, and that ice thickness and perennial ice coverage will decline in the central Arctic Ocean (Walsh *et al.*, 2005). The open water season may, therefore, be expected to increase in duration in the future (e.g., a decrease in maximum ice thickness, without accounting for changes in precipitation, may lengthen the open water season by 7.5 days per degree rise in temperature, Flato and Brown, 1996).

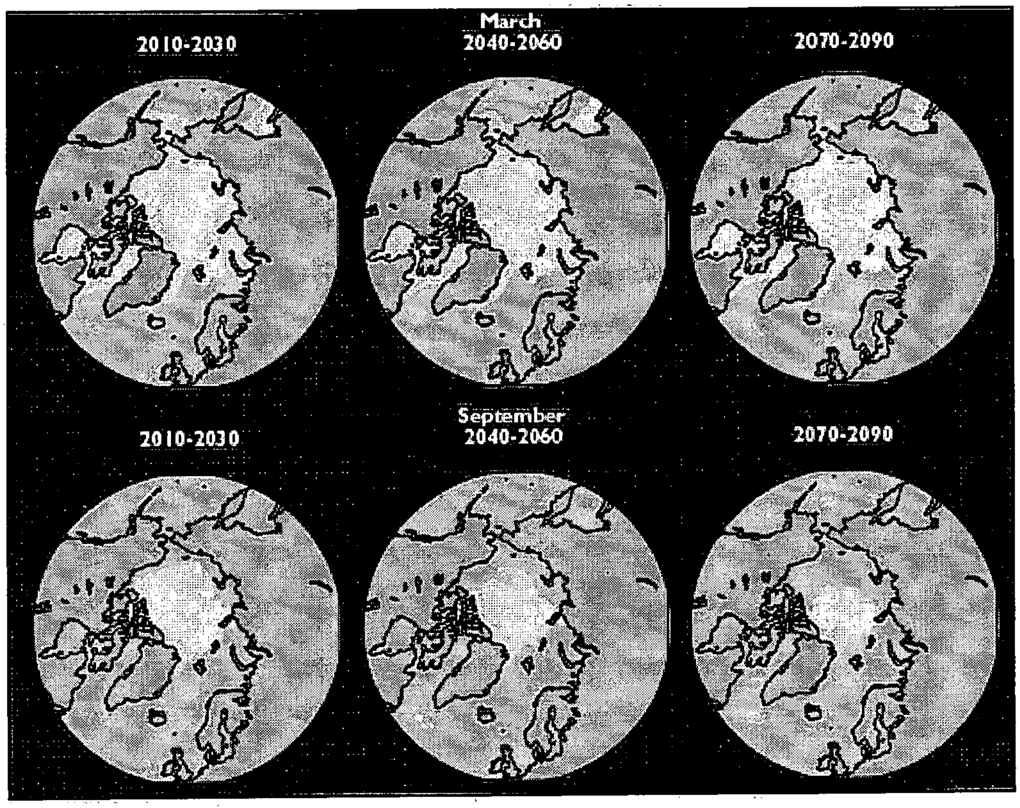


Figure 12. Projected change in extent of Northern Hemisphere sea ice to 2020, 2050 and 2080 (Walsh *et al.*, 2005, Figure 16.17). Sea ice extent is projected to change very little over the winter. The decline in fall ice extent is much more pronounced, and is evident along all coasts, with exception of northern Greenland. The decline in sea ice extent progresses through to 2080, with much of Nunavut, except the northernmost archipelago, free of ice at the end of the warm season.

Effects

Decline in sea ice extent under future climate forcing will affect the climate system, and the landscape and water resources of Nunavut (Walsh *et al.*, 2005). As sea ice declines in extent, arctic waters will warm over a longer open-water season, exacerbating loss of ice. As well, increased exposure of surface waters to rising temperatures will increase evaporative loss of water, increasing atmospheric moisture. This will likely increase cloudiness in coastal regions, which may offset rising temperatures and likely contribute to increased precipitation. Rising precipitation in coastal locations, along with increased exposure of coasts to wind, waves, storm surges and more dynamic, thin ice cover, will increase erosion of coasts, potentially displacing communities, damaging infrastructure and altering the quality of coastal freshwater bodies (Berner *et al.*, 2005). Thinner and less extensive ice will, however, benefit marine transport (e.g., reduce costs, increase access through the Northwest Passage), mineral extraction, and construction of offshore structures (Instanes *et al.*, 2005).

Glaciers and Ice Sheets

Background

Glaciers and ice sheets are prominent features in Arctic landscapes, useful indicators of climatic change, and important contributors of water to freshwater systems in the Arctic. Mass balances of glaciers and ice caps vary with temperature and precipitation, factors which affect the relative magnitudes of accumulation and ablation (Walsh *et al.*, 2005). Mass loss through melt generates runoff that is critical to some freshwater bodies of the Arctic. Glacial melt is the primary source of water for some rivers and lakes and, in others, is a fundamental part of the flow regime, sustaining high discharges well after the spring freshet has passed (Wrona *et al.*, 2005). Melt of glaciers and ice caps also influence sea levels, hence coastal processes.

Some glaciers and ice masses are more sensitive to climate change than others. For example, the mass balances of the most northerly ice masses in Nunavut (e.g., White Glacier and Devon Ice Cap) are nearly the least sensitive to temperature (-119 mm K^{-1} within circumarctic range of -49 to -831) and precipitation (3.7 mm $\%^{-1}$ within circumarctic range of 3.7 to 38.3) anomalies because of their highly continental climate and their exposure to temperatures that are for the most part near or below freezing. Glaciers of Baffin Island (i.e., Canadian Arctic >74°N) are slightly more sensitive (-190 and 5.9, respectively). The Greenland Ice Sheet is the least sensitive of all ice masses in the Arctic (-49 mm K^{-1}). As surface temperatures and snow accumulation change, so will the thermal regime, hence mass balance and sensitivity, of glaciers and ice caps.

Recent Trends

Land ice has a volume of approximately 3,1000,000 km^3 in the Arctic (Dowdeswell and Hagen, in press). Regionally, the magnitude of glacier and ice cap cover in Nunavut (151 800 km^2) is second only to that in Greenland (Walsh *et al.*, 2005). Ice masses here are dominated by highland ice, ice caps and outlet valley glaciers, some of which have tidewater termini. Those with a continental climate have large annual temperature range and abbreviated summer season, and are exposed to lower precipitation than glaciers in more maritime locations. Records from glaciers and ice caps of Nunavut indicate that mass balances changed over the past 40 years. Land ice experienced a net loss of mass from the 1960s to the mid 1980s (Koerner, 1996; Koerner and Lungaard, 1995), with mass loss increasing in subsequent years. This net reduction in mass of glaciers and ice caps of Nunavut has been attributed to rising temperatures and melt over the summer season, rather than to declining accumulation. Mass losses have been recorded in other regions of the Arctic as well, including western Canada (Koerner, 1996) and Alaska (Arendt *et al.*, 2002). This increase in melt of glaciers and ice caps in the Arctic over recent years has contributed, in small part (along with thermal expansion of sea water), to a 2 mmy^{-1} rise in global sea level over the 20^{th} century (Cabanes *et al.*, 2001), and a near 2.85 ± 0.42 mmy^{-1} increase over the 1990s (Walsh *et al.*, 2005). A similar rise in sea level has been documented for the Arctic (IPCC, 2001).

Future Changes

The ACIA projected changes in glaciers and ice sheets using a mass balance sensitivity approach, and expressed these changes in terms of sea level (Walsh *et al.*, 2005). Sensitivities of glacier mass balances were coupled with future temperature and precipitation as simulated by the 5 ACIA models, projecting mass balances and associated changes in future sea level. Projected increases in sea level with glacier and ice cap melt to 2100 vary between 0 and 6 cm for the various models. Regionally, this variability between models increases, and is highly limited in certainty. Van der Wal and Wild (2001) projected a change of 5.7 cm over a 70 year period, which is slightly higher than the model average of 4 cm rise in sea level to 2100. The IPCC (2001) projected an increase in global sea level between 11 and 77 cm between 1990-2100. Based on the ECHAM output (Figure 13), as forced by temperature alone, ice masses of the Canadian Arctic are projected to contribute to just slightly greater than 0.5 cm rise in sea level by 2100, much less than that of Alaska and Greenland.

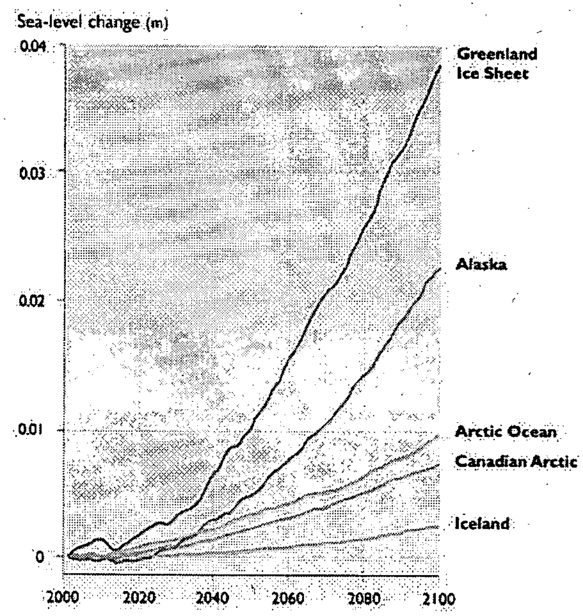


Figure 13. Change in sea-level through the 21st century, as projected by the ECHAM (Walsh *et al.*, 2005, Figure 6.20). Though glaciers and ice caps of arctic Canada are second only to Greenland in areal distribution, their projected contribution to sea level rise is much lower than that of Greenland and Alaska.

Effects

Increasing mass loss in glaciers and ice sheets over the 21st century will affect sea-level, as well as terrestrial and freshwater environments. Increasing sea-level has the potential to inundate coastal locations, particularly beaches and deltas, as well as other low-lying areas (i.e., areas with elevation < 10m above sea level, such as the western coast of Baffin Island and eastern coast of Nunavut along Hudson Bay, Walsh *et al.*, 2005). Sea-level rise, along with greater storm surges along ice-free coasts, will increase coastal erosion, particularly in coasts consisting of unlithified materials. Bedrock coasts of Nunavut are less sensitive to the effects of sea level rise than are coasts of western Canada (Figure 14), though beaches and deltas of the territory will erode quite readily. Sea-level rise will be further exacerbated by thermal expansion of sea water and coastal submergence (e.g., eastern Baffin Island, northwestern fringe of Arctic Archipelago), though offset by isostatic rebound in some locations of the Arctic (e.g., central Arctic and Hudson Bay) (Shaw *et al.*, 1994; Eggington and Andrews, 1989).

Increased mass loss with warming will alter discharges of melt water from glaciers and ice caps (Walsh *et al.*, 2005). This effect will vary regionally with thermal regime and precipitation. Rising temperatures will increase melt of ice masses, potentially resulting in net loss of mass. Enhanced melt will generate greater melt water flows from glaciers, affecting river discharges, riparian ecosystems and the formation of icings. Effects of increased land ice melt on northern communities may, therefore, include: impacts on hydropower generation and the availability of water for those communities that depend on glacier-fed lakes and reservoirs; impacts on coastal communities and infrastructure in low-lying areas as coastal erosion and potential flooding increase (e.g., coastal lowlands and tidewater glacier shores of Bylot, Devon and Ellesmere islands, Shaw *et al.*, 1998; IPCC, 1996); changes to iceberg production, affecting marine transportation, and; increasing salinity in estuaries and bays.

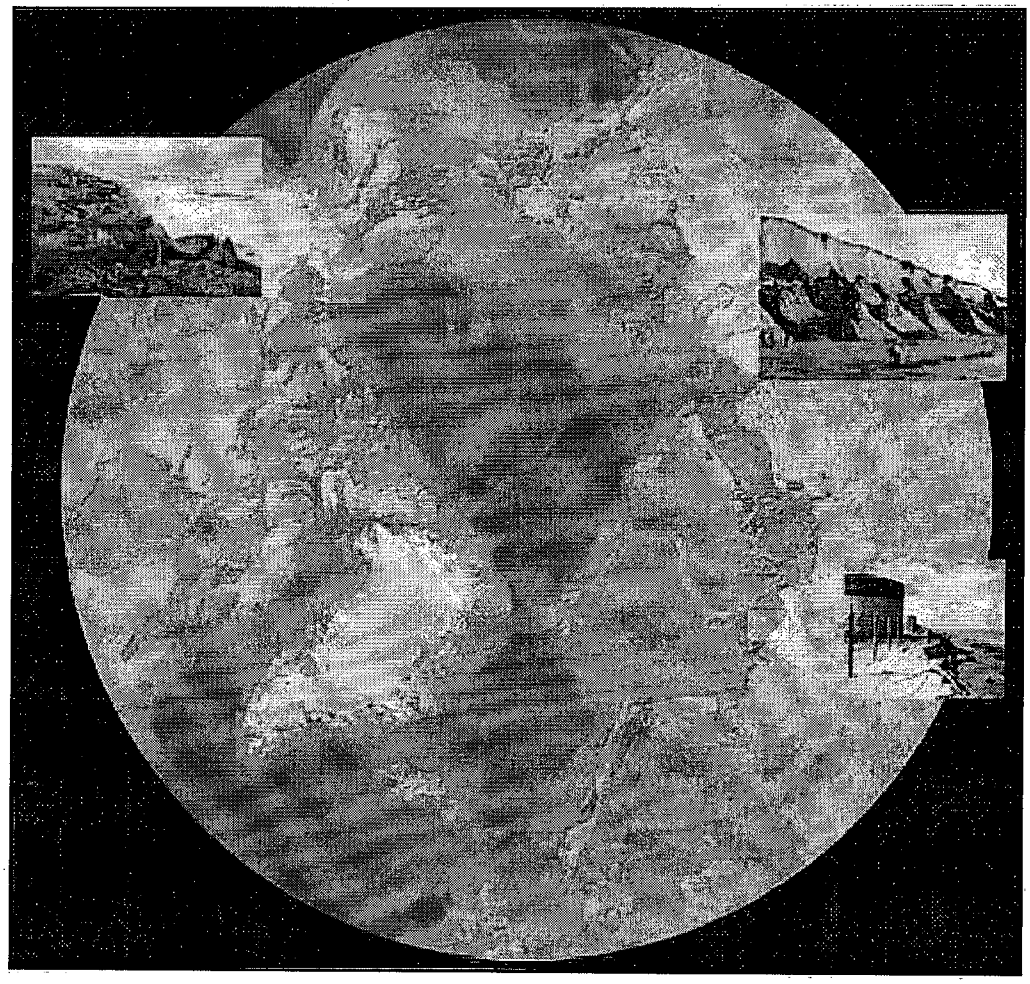


Figure 14. Coastal locations with elevation <10 m above sea level (red), and with lithified (brown) and unlithifed (green) materials (Walsh *et al.*, 2005, Figure 6.38). Coasts of Nunavut are sensitive to inundation due to sea level rise, however, they are less sensitive to coastal erosion than are unlithified coasts of western Canada and much of the Arctic.

Snow

Background

Annual and perennial snow are key to the hydrology of freshwater systems in the Arctic. The timing and magnitude of snow melt is highly variable regionally and inter-annually. Snow melt is the primary source of water for most lakes, ponds, rivers, streams and wetlands of the Arctic, and is a significant control on ice break-up and flooding in rivers. Changes to snow accumulation, storage and melt affect the "pulse" of water, nutrients, sediments and contaminants associated with high-latitude freshwater discharges in spring.

Snow is highly important to the thermal and moisture regimes of land and ice. Snow insulates, limiting ice development and permafrost thickness and depth. This effect is offset by the highly reflective nature of snow surfaces, which positively feeds back to cooling of the land/ice surface, and which delays spring melt (Groisman *et al.*, 1994). Snow cover affects, therefore, the development of the active layer, where most physical, chemical and biological soil processes take place, and limits the thickness of seasonal thaw. Snow also affects the moisture balance of land surfaces, with melt water contributing to soil water storage within active layer and wetland soils. Changes to snow cover will, therefore, have serious implications for the landscape and waters of the Arctic.

Recent Trends

Areal snow coverage in the Northern Hemisphere varies from <1 million km^2 in late August to 40-50 million km^2 in February (Figure 15). Overall, the areal coverage of snow in the Arctic has declined by near 10% from the 1970s to 2000 (David Robinson, 2003 in Walsh *et al.*, 2005), with a less pronounced decline in North America than Eurasia. This reduction in snow cover has been seasonally variable, occurring primarily in association with rising spring temperatures. Spring and summer snow covers over the last decade have been the lowest in the Northern Hemisphere over the 20th century (IPCC, 2001). Snow depths have also declined over recent decades in association with warmer temperatures (Brown and Braaten, 1998).

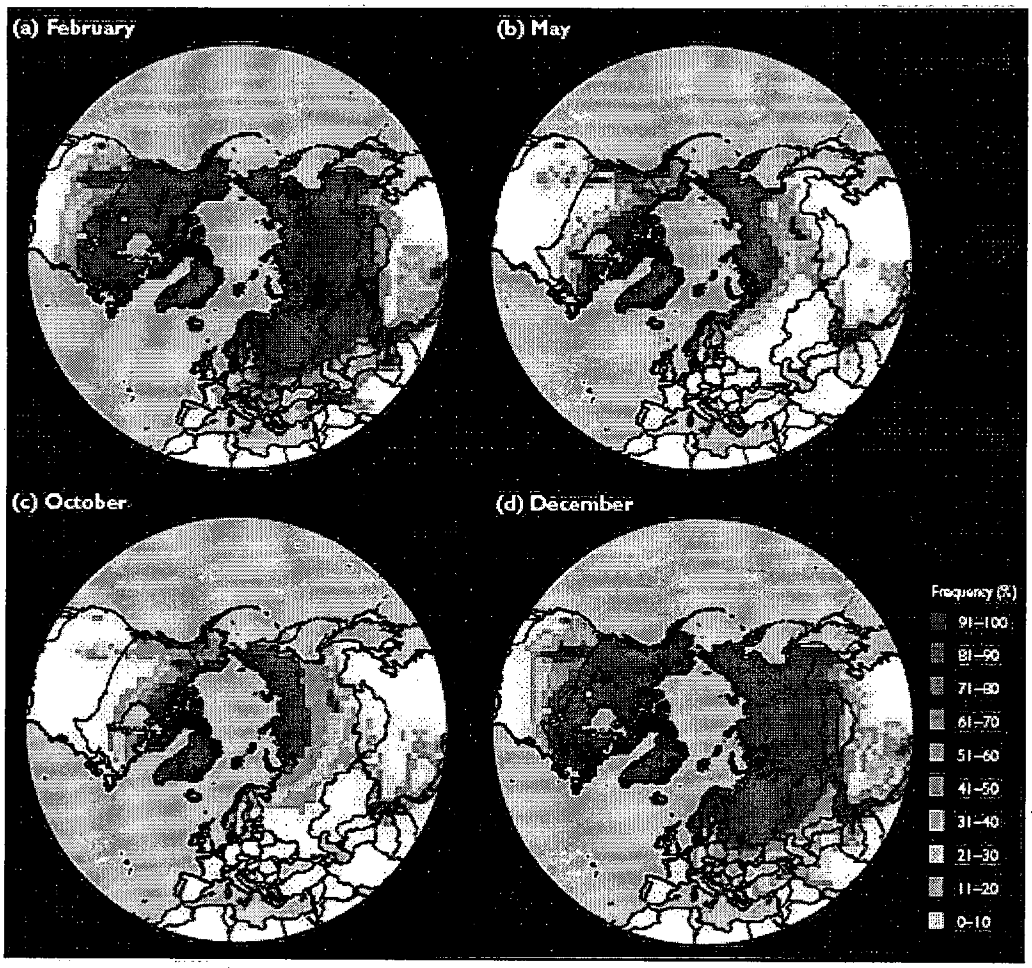


Figure 15. Frequency of snowcover. The mean position of the snow boundary is illustrated in green (Walsh *et al.*, 2005, Figure 6.11).

Future Changes

Snow cover is projected to decline in extent and duration over the 21st century. The 5-model mean change in annual snow cover is projected to range from -9 to -17% by 2070-2090 (Figure 16). This reduction will vary seasonally, with the greatest decline in snow-covered area projected for spring (April-May, model mean -4.9x10^6 km^2), followed by winter (-3.8), autumn (3.3), and summer (-1.1). Overall, snow distribution is projected to shift northward, and duration of the snow covered season to decline, as the length of the warm season extends (e.g., Figures 15 and 16). Snow depth and water equivalence are expected to increase at high latitudes and in more maritime locations as the frequency of stormy weather increases (Walsh *et al.*, 2005).

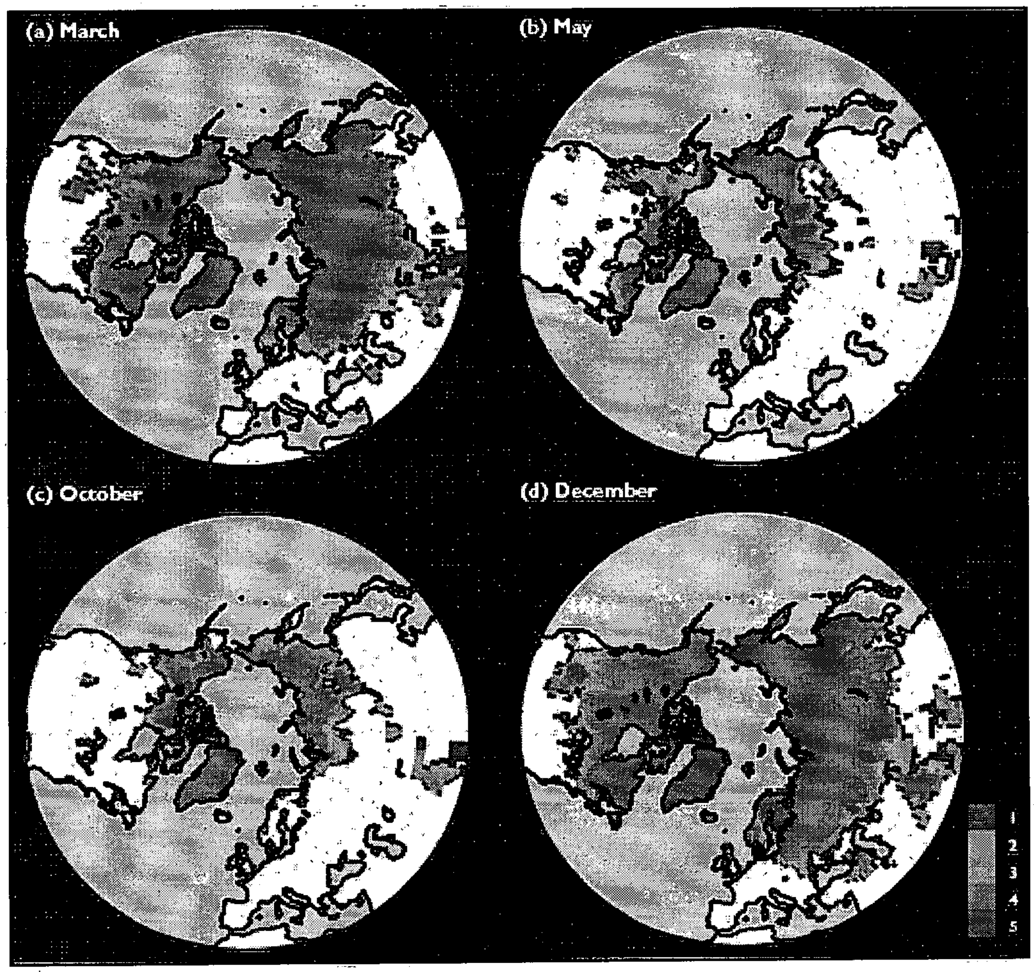


Figure 16. Seasonal distribution of snow cover projected for 2070-2090 (Walsh *et al.*, 2005, Figure 6.14). Colors indicate the number of models (out of the five used in the ACIA) that project snow for at least half of the years within the time slice. A comparison with the present-day frequency distributions (Figure 15) illustrates retreat of the southern snow boundary in Nunavut, particularly in May.

Effects

Reduced snow cover will affect soil temperature and moisture (Walsh *et al.*, 2005). Reduction in the length of the snow-covered season with earlier and more pronounced spring warming, and potentially later autumn cooling, will result in warmer soils, particularly over the warm season. Potentially deeper snow packs, resulting from the projected increase in winter precipitation for Nunavut, may further moderate winter soil temperatures. The opposite may be said for locations projected to experience a reduction in snowfall. As such, the depth of seasonal thaw may be expected to increase through much of the Arctic, and the thickness and distribution of permafrost may be expected to decline, as snow extent and duration change. This, in turn, will potentially result in decreased soil stability, increased erosion, altered water storage in catchments, reduced quantity and quality of runoff, and increased biogeochemical cycling and production of greenhouse gases (which may positively feedback to climate change, Callaghan *et al.*, 2005).

Changes to snow cover will also have serious implications for high latitude freshwater hydrology (Wrona *et al.*, 2005). The projected increase in snow depth and decline in snow duration and extent will potentially alter the composition of ice (e.g., white ice, which includes snow, is more reflective than clear ice) and reduce ice thickness and duration, thereby affecting the length of the open-water season. River flows, which are typically nival at high latitudes and characterized by dramatic ice-flow dynamics with spring freshet, will likely shift in magnitude and timing annually, and experience less pronounced ice break-up and flooding in spring with earlier melt and reduced ice thickness. River channel morphology will, therefore, be affected as well, as the magnitude of scour and floodplain development decline with potentially reduced flood discharges.

Permafrost

Background

Permafrost is widespread through the Arctic and is intrinsic to the physical nature of high latitude terrestrial and aquatic landscapes, as well as their hydrology and biology. Permafrost consists of terrestrial matter (e.g., soil, rock, sediment) that has remained at a temperature of 0°C or less for a period of two or more years (Walsh *et al.*, 2005). Permafrost may have a high water content. Such "ice-rich" permafrost is more susceptible to dramatic change in physical form with altered thermal regime than is permafrost that is not ice-rich. Permafrost consisting of unlithified materials will also be highly susceptible to modification under future climates, as will that existing near the melting point (i.e., within 1 to 2°C, and several meters thick). Interactions between climate and the surface of colder permafrost (i.e., -10° or less, and 500 to 1400 m thick), such as that found throughout much of Nunavut, are manifested primarily in the seasonal thaw layer. This 'active layer' may extend to depths from several tens of centimeters down to a couple meters over the warm season, refreezing over the cold season.

The depth to which the active layer thaws, as well as the distribution and depth to which permafrost develops, depend on many factors, including the magnitude and duration of cold/warm season temperatures, and the magnitude and composition of precipitation. Warmer permafrost and thicker active layers are associated with warmer air temperatures, deeper (insulating) snow packs, and lower albedo surfaces (e.g., vegetation, bare ground). Permafrost warming is moderated by such things as organic soils, which have low thermal conductivity, and latent heat exchanges associated with refreezing and evaporation. The active layer is of prime importance to the biogeochemistry and hydrology of terrestrial surfaces and aquatic environments in a number of ways, including: subsurface biogeochemical processes such as decomposition; establishment of vegetation; quality and quantity of runoff to lakes, ponds and rivers; water levels; nature and distribution of wetlands and drainage networks.

Recent Trends

Permafrost underlies approximately 24% of the terrestrial Northern Hemisphere (Zhang *et al.*, 1999) and occurs in continuous, discontinuous and sporadic distribution throughout the Arctic (Figure 17). Permafrost of Nunavut is continuous (90 to 100% areal coverage), with mean annual temperatures near the base of the seasonal thaw layer varying from near -2°C at lower latitudes to near -19°C at higher latitudes, and with thicknesses reaching up to 600 m (The National Atlas of Canada, 1995). Ground ice represents approximately 10% of the upper 10 to 20 cm of permafrost found through much of central Nunavut and Baffin Island, and varies from <10% to isolated areas of >20% in the Archipelago. Recent changes in permafrost in the Arctic have been spatially variable, with the greatest warming occurring in Alaska (Nelson, 2003; Lachenbruch and Marshall, 1986), and the least in Siberia and eastern Canada. Permafrost warming in northeastern Canada has been slightly more pronounced through the 1990s (Allard *et al.*, 1995; Brown *et al*, 2000), with a $0.15°Cy^{-1}$ rise in permafrost temperatures in Alert, Nunavut (1995-2000; Geological Survey of Canada in Walsh *et al.*, 2005).

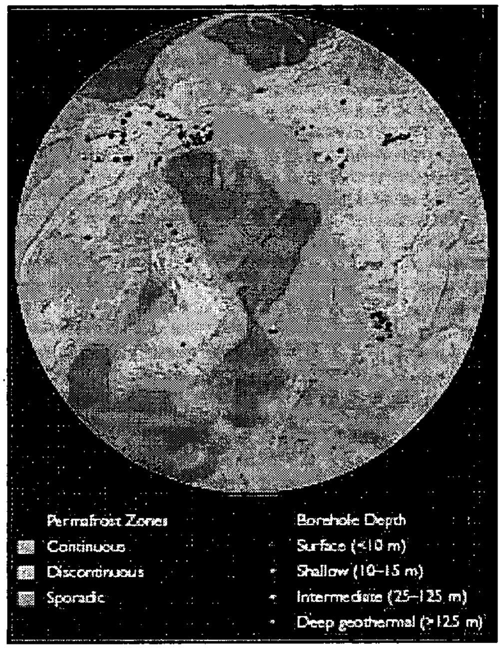


Figure 17. Distribution of continuous, discontinuous and continuous permafrost in the Arctic (from Romanovsky *et al*., 2002 in Walsh *et al.*, 2005, Figure 6.21). Circles are proposed locations of boreholes for permafrost temperature monitoring by the Global Terrestrial Network for Permafrost (GTN-P).

Future Changes

Projected change in areal coverage of circumarctic permafrost to 2030, 2050 and 2080 has been determined using a frost-index-based model driven by the 5 ACIA GCMs (Walsh *et al.*, 2005; Anisimov and Nelson, 1997). According to the median model (GFDL), circumarctic permafrost will lose 23% of its current area by 2080, with continuous permafrost declining by 18% to 2030, 29% to 2050, and 41% to 2080, resulting not only in loss of permafrost, but also in a likely northward shift in the permafrost zones. Generally, the distribution of continuous permafrost in Nunavut is expected to remain stable to 2080 (Figure 18), particularly in the Archipelago.

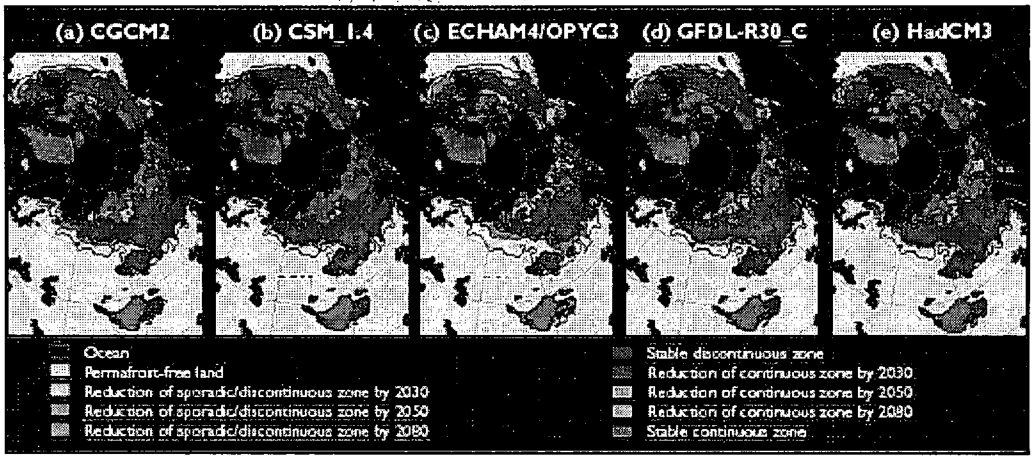


Figure 18. Projected changes in circumarctic permafrost distribution (Walsh *et al.*, 2005, Figure 6.23). The permafrost boundary is projected to shift northward resulting in contraction of the continuous permafrost zone and repositioning of the discontinuous and sporadic permafrost zones to higher latitudes. Coupling of the frost-index and ECHAM models indicates little potential change in the areal distribution of permafrost in northern Nunavut, and contraction of the continuous permafrost zone within south-central Nunavut.

Region-specific projections of future changes in permafrost have been developed through coupled GCM-soil models (Zhang *et al.*, 2001). Mean annual ground temperature of Alaskan permafrost is projected to increase by up to 5°C by 2100, with active layer thickness increasing by up to 1 m. Active layer depth is projected to double in Siberia, increasing from < 40 cm at present to 60-100 cm on average by 2100. Projections of future seasonal thaw depths were also made based on the ACIA GCMs and assumptions regarding soil, organic layer, vegetation and snow cover (Anisimov *et al.*, 1997). Active layer depth was projected to increase by > 50% at high latitudes in the Arctic by the middle of the 20th century, and by 30 to 50% in other areas (Figure 19). For example, Canada's arctic coast is expected to experience a +2-3°C change in air temperature, a +2-2.5°C change in permafrost temperature, and a 50% increase in depth of seasonal thaw (Walsh *et al.*, 2005). The depth of the active layer in Nunavut is, for the most part, expected to increase by > 50% by 2050, much in the same manner as other coastal areas in the Arctic, with a 30-50% increase in active layer depth at more southerly latitudes within the territory.

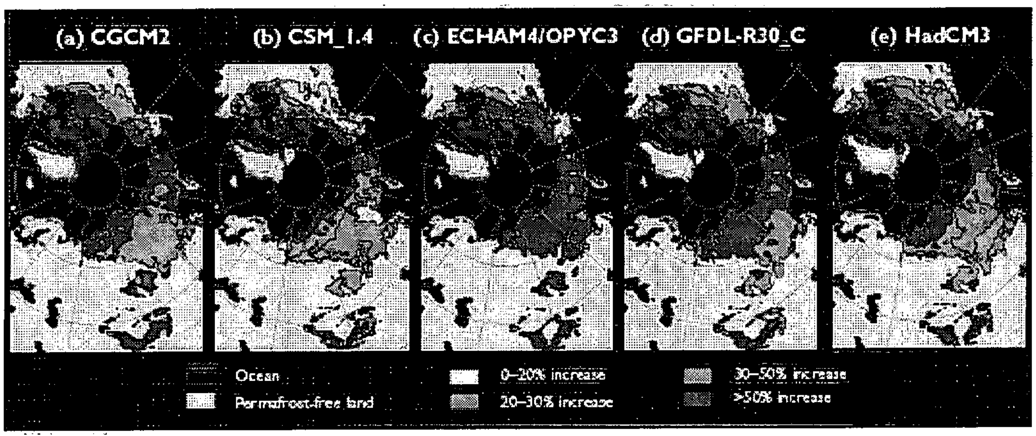


Figure 19. Changes in circumarctic active layer depth to 2050 as projected by the 5 ACIA GCMs (Walsh *et al.*, 2005, Figure 6.24). Increases in depth of the active layer are projected to be greatest at higher latitudes in North America and Siberia. Active layer depths in Nunavut are expected to increase by > 50%. Local changes in active layer will vary with such things as topography, hydrology, vegetation and human activity.

Effects

The greatest impacts of permafrost degradation in Nunavut will be associated with increasing thickness of the active layer. Changes to the active layer will alter the moisture regime of terrestrial surfaces and will ultimately affect the quantity and quality of water in freshwater bodies. As the thickness of the active layer increases, the infiltration and storage capacity of terrestrial surfaces will increase, likely reducing surface runoff and increasing groundwater discharges (Walsh *et al.*, 2005; Wrona *et al.*, 2005). Catchment flows to rivers, lakes, ponds and wetlands may, therefore, be expected to decline, particularly in spring when a deeper active layer and greater infiltration will reduce the magnitude of spring runoff and increase groundwater contributions later in the season. This effect may be offset by projected increases in snowfall (Figure 5), which will likely enhance overall moisture availability in the Arctic, or may be exacerbated by deepening of the active layer with increased snow pack depth and earlier snow melt. Permafrost melt, in altering the moisture balance of land surfaces and in increasing thermokarst activity (i.e., melt of ground ice and subsequent subsidence), will alter the distribution of drainage networks, ponds and wetlands, promoting development of new aquatic systems and resulting in drainage of others (e.g., Marsh and Neumann, 2001; Osterkamp *et al.*, 2000; Fedorov, 1996; Mackay, 1992). Increased drainage with permafrost melt and increased evaporation under warmer climate will result in drying of some freshwater systems, such as unglaciated lowlands of the Archipelago and associated areas of ponded water (e.g., Polar Bear Pass on Bathurst Island, Truelove Lowland on Devon Island), which are particularly sensitive to water deficit (Wrona *et al.*, 2005).

Future increases in surface moisture will not only affect runoff, but will also, along with rising temperature of permafrost soils, have implications for terrestrial and aquatic biogeochemistry (Walsh *et al.*, 2005; Callaghan *et al.*, 2005; Wrona *et al.*, 2005). Rising temperatures will increase microbial activity in permafrost soils and, along with greater soil moisture, will enhance the generation and release of carbon-based greenhouse gases, which may positively feedback to climate warming (e.g., methane production is greater in warmer, more moist soils, Friborg *et al.*, 1997; Funk *et al.*, 1994). Further releases of greenhouse gases to the atmosphere will likely result from melt of permafrost, which has been shown to be a substantial perennial store of methane (Panikov and Dedysh, 2000). Initial high magnitude fluxes of greenhouse gases from degraded permafrost landscapes will stabilize over time (e.g., Turetsky *et al.*, 2002, 2000; Camill, 1999) and, with likely increased vegetative growth (Callaghan *et al.*, 2005), will be offset by increased photosynthetic uptake of carbon dioxide. More abundant and productive vegetation growth on warmer, permafrost-degraded soils will also increase the availability of carbon and nutrients in soils and runoff. Water quality may, therefore, deteriorate with increased nutrient and sediment loading from catchments.

Increasing depth of the active layer and ground ice melt with rising temperatures will reduce the stability, and increase erosion, of permafrost soils (e.g., Weller and Lange, 1999). Shallow landslides (active layer failures) have been documented on Fosheim Peninsula, Ellesmere Island, in association with higher temperatures and precipitation (Lewkowicz, 1990). Thaw of ground ice on Banks Island resulted in the development of hummocks and depressions that took 15 years to stabilize (Maxwell, 1997). Increased erosion, subsidence and mass movements will, therefore, likely threaten the integrity of infrastructure in Nunavut, though less so than in areas of the Arctic where permafrost is projected to disappear entirely and where unlithified materials predominate (Walsh *et al.*, 2005; Instanes *et al.*, 2005). Though potential damage to infrastructure will be detrimental to northern communities, warmer soils and increased soil moisture will promote economic development, increasing land use activities such as agriculture and mining in some areas (Walsh *et al.*, 2005).

River and Lake Ice

Background

River and lake ice are prominent features in the Arctic, and exert significant control on both aquatic and terrestrial systems. Freshwater ice affects the thermal and radiative regimes of lakes, ponds, rivers, streams and wetlands (Wrona *et al.*, 2005). Ice covers limit water temperatures, influence biological productivity (e.g., abbreviated growing season, low dissolved oxygen and radiation receipt over cold season, reduced carbon and contaminant processing), and affect the nature and fate of freshwater discharges. Ice is a critical component of river discharge regimes in the Arctic, affecting the timing of break-up and the magnitude of flooding and disturbance associated with spring freshet. River and lake ice affect, therefore, the transfer of water, biota, sediment, carbon, nutrients and contaminants to receiving water bodies, and to riparian, floodplain and delta lakes, ponds and wetlands (e.g., Prowse and Conly, 2001; Lesack *et al.*, 1991; Marsh and Hey, 1989).

Freeze-up of arctic lake and river ice typically occurs earlier, and break-up later, at higher latitudes (e.g., Kratz *et al.*, 2000; Assel and Herche, 1998). Timing of freeze-up is highly dependent on the depth and heating of water over the preceding warm season (Vavrus *et al.*, 1996; Wynn *et al.*, 1996). Timing of break-up is dependent on antecedent energy conditions (e.g., Anderson *et al.*, 1996; Assel and Robertson, 1995; Robertson *et al.*, 1992; Palecki and Barry, 1986). The magnitude of break-up-associated effects on hydrological, physical and biological environments of the Arctic is not only affected by the thermal regime of ice and water, but also by latitudinal variation in catchment snow pack and river ice melt. This is particularly true for large rivers, in which snowmelt runoff generated at lower latitudes breaks up ice at higher latitudes, resulting in ice-jams and associated flooding (Prowse, 2001; Prowse and Gridley, 1993). Walsh *et al.* (2005) and Wrona *et al.* (2005) discuss these topics in further detail.

Recent Trends

The timing of lake and river ice break-up and freeze-up have changed over the past century. From 1846 to 1995 there was a trend toward earlier break-up and later freeze-up in lakes and rivers globally (Magnuson *et al.*, 2000). A similar trend was found in rivers of the Former Soviet Union and western Siberia from the late 1800s to 1985; river ice freeze-up was delayed and break-up was advanced by 7-10 days over a 100 year period (Soldatova, 1993; Ginzburg *et al.*, 1992). The opposite was documented, however, in rivers of central and eastern Siberia. Break-up has also advanced in lakes of southern Finland (Kuusisto and Elo, 2000) and rivers of northern Sweden, Finland and Latvia (Kuusisto and Elo, 2000; Zachrisson, 1989).

In North America, average ice break-up of rivers in the northwest has occurred nearly 5 days early over the last 100 years (Sagarin and Micheli, 2001; Jasek, 1998). This is consistent with the findings of Zhang *et al.* (2001) that indicate an advance in the break-up of river ice in western Canada over the past 50 years, and with paleoclimatic data that indicate a longer-open water season in Canada's Arctic (e.g., Michelutti *et al.*, 2002). Though spring melt has been observed to occur earlier in Nunavut (Krupnik and Jolly, 2002), delays in spring melt have also been noted (e.g., eastern Hudson Bay, McDonald *et al.*, 1997). Unlike in many other areas of the Arctic, river ice freeze-up in Canada has generally occurred earlier over recent decades.

Future Changes

Future changes in ice break-up and freeze-up may be related to projected spring and autumn air temperatures (Soldatova, 1993; Ginzburg *et al.*, 1992). The use of air temperature as a proxy for ice dynamics has limited accuracy (Bonsal and Prowse, 2003; Livingstone, 1999). Nevertheless, general conclusions may be made regarding potential future changes in river and lake ice regimes using simulated air temperatures and taking into consideration the potential effects of altered heat fluxes, timing and magnitude of precipitation, and catchment runoff (e.g., Walsh *et al.*, 2005; Bonsal and Prowse, 2003; Prowse and Beltaos, 2002).

Freshwater ice is projected to decline in thickness and duration, altering the hydrology of arctic freshwaters. The length of the ice-covered season will increase as higher spring air and water temperatures occur earlier and as low autumn temperatures occur later. This lengthening of the ice-free season (i.e., period of heating) and shortening of the cold season (i.e., period of ice growth) will likely result in a decline in lake and river ice thickness (Walsh *et al.*, 2005). Ice thickness and duration will also decline as snow pack depth increases (Prowse and Gridley, 1993), though snow accumulations may reflect radiant energy and slow ice decay, thereby extending the ice-covered season.

Effects

Projected changes to ice will affect water quantity and quality in freshwater systems (Wrona *et al.*, 2005). As the length of the open-water season extends, and air temperatures and potential evaporation increase, evaporative water loss from lakes, rivers and wetlands will increase. This may result in declining water levels, hence drying out of some systems over the warm season. Water losses will likely be less pronounced in coastal locations, which are projected to experience an increase in precipitation. Winter water levels may be expected to increase in some systems as ice covers thin and as baseflow from permafrost degraded catchments increases. Lowering of water levels and lengthening of the open-water season in some systems will not only have implications for water availability in the Arctic, but also for water quality (see following section for further discussion).

The most dramatic effects of decreased ice thickness and duration will be associated with alteration of river ice break-up and associated hydrology. As ice break-up advances in timing, ice thickness declines, and latitudinal temperature gradients dampen, break-up and associated flooding will likely decline in severity (Gray and Prowse, 1993; Prowse and Beltaos, 2002). This less dynamic (more thermal) ice break-up, will alter the disturbance regime of rivers and decrease the transfer of water and associated loadings to riparian areas and floodplains (Prowse and Culp, 2003; Cunjak *et al.*, 1998; Scrimgeour *et al.*, 1994). This effect will be particularly pronounced in large rivers, which receive substantial discharges of spring melt water from large catchments, and which may be exposed to less pronounced latitudinal gradients in temperature under future climate. Increased winter precipitation and associated snowmelt runoff may, however, increase or sustain the intensity of break-up and flooding in some areas.

Changes to ice quality (e.g., mechanical strength) and quantity may have serious implications for transportation and infrastructure (Instanes *et al.*, 2005). Winter ice-cover routes may be disrupted, and acceptable load size and weight will likely decline. Changing ice regimes will affect shipping, potentially reducing costs with longer ice-free season on rivers. Lowered water levels may, however, negate this benefit, restricting transport by river. Hydroelectric operations will need to adjust peak generating capacity and ensure sustainable management of water resources.

Water Quality

Background

The primary sources of water for communities in the Arctic are lakes, ponds, rivers, streams and impoundments (Berner *et al.*, 2005). Groundwater sources are restricted and of lower quality than surface waters. Freshwater bodies are fed primarily by snow and ice melt, hence are vulnerable to changes in precipitation (e.g., quantity and timing), melt (e.g., perennial snow and ice), permafrost (e.g., catchment characteristics), river flow (e.g., freshet), and lake ice (e.g., length of the ice-free season) (Diamond *et al.*, 2003; Macdonald *et al.*, 2000; Braune *et al.*, 1999; Kidd *et al.*, 1995). Surface and ground water quality may be degraded by contaminants (e.g., persistent organic pollutants, polycyclic aromatic hydrocarbons, mercury, Wrona *et al.*, 2005; AMAP, 1998), sediment loading (e.g., turbidity), and nutrient and organic carbon loading (e.g., eutrophication). Once in the hydrologic cycle, contaminants, sediments and nutrients may be modified through biogeochemical cycling (e.g., carbon cycling), food webs (e.g., bioaccumulation) and temperature-contaminant interactions (e.g., revolatilization). This report addresses the potential impacts of future climate change on the physical and chemical quality, hence potability, of water resources in Nunavut. More thorough discussion of water quality issues in northern environments are found in the ACIA (2005) and AMAP (1998, 2003).

Future Changes and Effects

Water quality in Nunavut will change as climate change progresses. Future water quality will be affected by the projected increase in precipitation, melt of perennial stores of ice and snow, melt and erosion of permafrost catchments and associated loadings to aquatic systems, potential decreases in runoff and water levels with increasing thickness of the seasonal thaw layer, less dramatic spring freshet and associated flooding, potential damage to infrastructure, sea level rise, and human activity. Generally, water quality may be expected to decline as these changes progress, with likely increased contaminant, sediment and nutrient loadings, increased intrusions of coastal marine waters and contamination from anthropogenic sources.

Increases in precipitation and more pronounced melt will increase the risk of water quality degradation in the Arctic under wetter, warmer climate over the 21st century. As temperatures rise, the ability of the atmosphere to transport moisture from low latitudes to high latitudes will increase (Serreze *et al.*, 2000; Manabe *et al.*, 1992). Increased precipitation may amplify atmospheric deposition of contaminants and particulates, and significantly increase capture of hydrophilic (i.e., partition strongly into water) contaminants in the Arctic (Wrona *et al.*, 2005). As temperatures rise and annual and perennial snowpacks melt, contaminants such as persistent organic pollutants may vaporize or be redistributed to soil, vegetation and melt water. Melt of perennial stores of snow, as well as ice and permafrost, may result in the release of substantive, episodic contaminant loadings to freshwater bodies (i.e., "shock loading", Blais *et al.*, 1998, 2001).

Increasing thickness of the active layer will also affect water quality in Nunavut, altering loadings to lakes, ponds, streams and rivers (Walsh *et al.*, 2005; Wrona *et al.*, 2005). Permafrost is a significant reserve of sediments, nutrients and organic carbon, and is a limiting factor in the abundance and productivity of vegetation in the Arctic. As the thickness of the active layer increases and permafrost melt and erosion progress, sediment, nutrient and organic carbon loads in runoff will increase, though perhaps more so in association with groundwater than surface runoff. Transfer of these loads to lakes, ponds and wetlands via rivers and streams, and newly developed dendritic drainage networks (e.g., McNamara *et al.*, 1999), will increase turbidity and eutrophication (i.e., increase in primary productivity with reduced temperature- and nutrient-limitation, Wrona *et al.*, 2005). Loadings of dissolved organic carbon from permafrost-degraded catchments will increase further with the projected rise in primary productivity and establishment of woody vegetation on warmer soils over a lengthened growing season (Callaghan *et al.*, 2005; Vincent and Hobbie, 2000; Fallu and Pienitz, 1999; Rühland and Smol, 1998; Pienitz and Smol, 1993). Changes to vegetation will also affect contaminant cycling in the Arctic, with leafy plants "pumping" persistent organic pollutants from the atmosphere. This effect, along with the capacity for soil organic carbon to store contaminants, will increase contaminant capture and storage in terrestrial and freshwater systems of the Arctic (Wania and McLachlan, 2001; Simonich and Hites, 1994).

Changes to flow and ice regimes in rivers and lakes, and altered water balances in wetlands, will also have implications for water quality in the future (Walsh *et al.*, 2005;

Wrona *et al.*, 2005). Rivers act as conduits of contaminants from arctic and extra-arctic catchments to lakes, ponds and wetlands. As temperature and precipitation increase, river runoff and contaminant transfer northward from southern industrial and agricultural areas will increase. Runoff and associated contaminant burdens will increase in some wholly arctic catchments as well. This effect will be moderated by the projected increase in active layer depth in Nunavut and the associated increase in infiltration and contaminant capture within catchment soils. Those contaminants that are successfully transported to receiving waters, particularly in association with increased organic carbon and sediment loads from eroding, permafrost-degraded catchments, will be more effectively incorporated into the water and sediment of these systems as the length of the ice-free season, mixing, and primary productivity increase (e.g., Diamond *et al.*, 2003; Macdonald *et al.*, 2000). Lakes and ponds, which under current climate are effective stores of sediment and associated contaminants, may become greater sources of contaminants to the atmosphere (i.e., net evasion of contaminants from warmer, saturated waters, Harner 1997) and to the food web (i.e., increased biogeochemical cycling and bioaccumulation of contaminants as temperatures and light penetration increase). Lowering of water levels due to higher temperatures, longer open-water season and greater evaporation, will increase contaminant concentrations, hence potential toxicity, in some systems. This effect will be negated in part by increased precipitation and cloudiness, particularly in coastal locations. Wetlands also have the potential to contribute contaminants to freshwaters. Alteration of water levels in wetlands has serious implications for the production of methyl mercury, which is highly hazardous to the health, and which may form in newly formed or flooded wetlands (Suchanek *et al.*, 2000; Driscoll *et al.*, 1998; Bodaly *et al.*, 1984). As well, changes to permafrost, along with erosion and combustion of peat-rich soils, will increase loading of polycyclic aromatic hydrocarbons to freshwaters (Yunker *et al.*, 1993, 2002).

Changes to hydrology and permafrost further have the potential to degrade water quality through damage to water distribution and sanitation infrastructure, potentially increasing contamination of water supplies in northern communities (Berner *et al.*, 2005; Instanes *et al.*, 2005). A more detailed discussion on the effects of permafrost degradation on infrastructure is given in the following section. In the Arctic, water is distributed by community or piped systems (above- or below-ground), and wastewater and solid wastes are disposed of in community receptacles, lagoons, tundra ponds, pit privies, holding tanks or septic systems. As temperatures rise, permafrost soils will lose stability and erode, particularly in areas that are ice-rich. Shifts in permafrost have the potential to degrade water supplies not only through damage to the distribution system itself, increasing the vulnerability of water supplies to contaminant exposure, but also through damage to sanitation infrastructure, which may result in spread of wastes and transmission of disease (IHS, 1999; Schliessmann *et al.*, 1958). This risk of contamination will increase with flooding from storm surges and increased river discharges, heavy rainfall, and thaw of waste pits, which may result in degradation of local water supplies as well as those further downstream (Harris, 1987). Changes to permafrost may also result in restricted access to water supplies (Geldreich, 1992). For example, erosion of warming soils and loss of portions of Sarichef Island (Chukchi Sea) have limited access of members of an Inupiat Eskimo village to water and wastewater haul systems (Berner *et al.*, 2005). Though rising temperatures have the potential to adversely affect water supplies in the future, warmer conditions will reduce the risk of flooding and associated damage in some locations, and will increase the efficiency of biological treatment of wastes (Berner *et al.*, 2005).

Future increases in sea level may threaten the integrity of coastal water supplies. Salinity, dissolved solids and contaminants in coastal water supplies may increase as coastal storminess and erosion progress (Linsley *et al.*, 1992). Seawater contamination of water supplies can make water non-potable; bromide has the potential to increase the formation of carcinogenic disinfection byproducts (Singer, 1999), the treatment of which is highly complex and costly.

Finally, contaminant loading to freshwaters will likely increase as human activity at high latitudes increases. As climate at higher latitudes becomes more moderated, with warmer temperatures, longer growing season and less coverage of snow, ice and frozen-soils, accessibility and operating conditions will improve. Agricultural and industrial practices and associated contaminant loadings will, therefore, likely increase. Steps will have to be taken to mediate the potential effects of increased loadings of pesticides and herbicides to freshwater systems, and of contaminants to the atmosphere (Wrona *et al.*, 2005; AMAP, 1998).

Infrastructure

Background

The design, construction and maintenance of infrastructure meet unique challenges in the Arctic relative to locations with more temperate climate. Permafrost is paramount as a foundation for infrastructure including community buildings, roads and extraction facilities (Instanes *et al.*, 2005). The depth of the active layer and ground temperature must be considered in the design of infrastructure, as they control processes such as creep, adfreeze bond, thaw settlement, and frost heave and jacking.

Future Changes and Effects

Infrastructure will be adversely affected as climate change progresses in Nunavut. The most significant impacts to infrastructure will be associated with increased thickness of the active layer, altered flow and ice regimes, increased coastal erosion, and increased occurrence of extreme events. Projected changes to active layer thickness and precipitation will not only potentially cause damage to present-day infrastructure, but also affect the design of future structures (e.g., Dai and Wang, 1995; Nixon, 1990a; Nelson *et al.*, 1983; Johnston, 1981).

Projected increases in active layer thickness may be related to air freezing and thawing indices, which are used in the design of permafrost engineering (Instanes, 2003). The magnitude of the thaw season (i.e., thickness to which the active layer will develop) may be determined from air thawing and freezing indices (ATI, AFI), which are derived from annual variation in air temperatures that are greater than and less than 0°C, respectively. High ATIs are selected in the design of more sensitive structures, such as pipelines, to account for maximum potential thaw (e.g., once in 1000 years probability of occurrence). Figure 20 illustrates historical and projected freezing and thawing indices for Kugluktuk (Coppermine), Nunavut, from 1940 to 2100. The AFI declined over the 20th century, and is projected to decline further through the 21st century in association with a 5-6°C increase in winter temperatures from 1960 to 2100 (Instanes *et al.*, 2005). The ATI is projected to increase through the 21st century, equivalent to a 4°C increase in summer temperatures from 1960 to 2100. These results are consistent with the projected increases in warm and cold season temperatures for the Arctic.

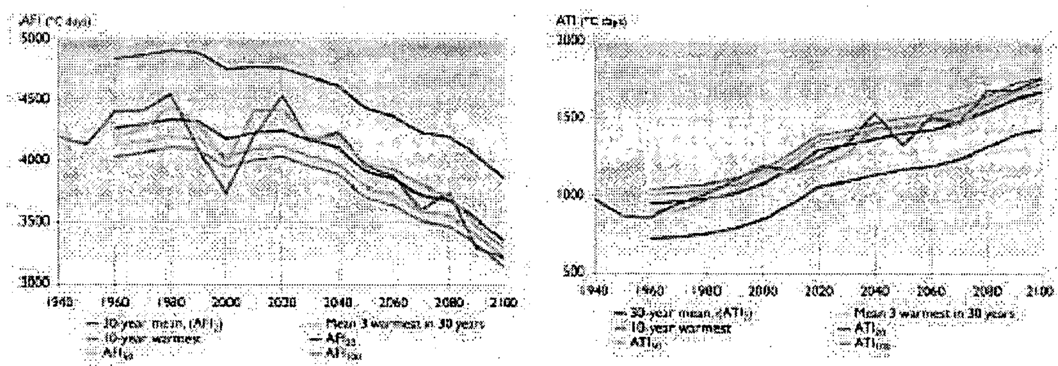


Figure 20. Change in freezing and thawing indices for Kugluktuk (Coppermine), Nunavut, from 1940 to 2100 (Instanes *et al.*, 2005, Figures 16.26, 16.27). The solid black line is the 30-year mean.

As the length of the thaw season and the thickness of the active layer increase, mean ground temperatures will increase, decreasing soil strength (Table 8). Soil strength is projected to decline by between 23 and 48% in select locations of northwestern North America to 2090-2099, and by between 12 and 34% for locations within Russia and Siberia. As physical and mechanical properties of soils change with rising temperatures, the following effects may be expected: increased creep and loss of adfreeze bond with warming at depth; thaw settlement and increased frost-heave with increasing depth of the active layer; decreased pilings length, and increased landslides and surface settlements with development of a residual thaw zone (talik) (Esch and Osterkamp,1990).

Table 8. Projected changes in mean ground temperature and soil bearing strength in northwestern North America, Russia and Siberia to 2090-2099 (Instanes *et al.*, 2005). Values are for depth of 10 m. Projected loss in strength are greatest for North America, and imply that damages to infrastructure may be greatest in locations such as these.

Site	**Mean temperature 1990-1999 [°C]**	**Mean temperature 2090-2099 [°C]**	***Soil strength loss in %**
Barrow	-12	-7	23
Bethel	-2	0	40
Fort Smith	-4	0	48
Naryan Mar	-4	-3	12
Nuuk	-2.5	-0.5	34
Svalbard Airport	-6	-4 (2040-2049)	17
Turukhansk	-7	-5	15
Verkhoyansk	-16	-12	14

* calculated after Ladanyi, 1996.

As the active layer increases in thickness in Nunavut, and soil bearing strength declines, the susceptibility of engineered structures to geocryological hazards will increase (Figure 21, Instanes *et al.*, 2005). Potential for thaw-induced settlement is projected to be primarily low in North America and high in Eurasia by 2050, with low to moderate risk in Siberia. Risk of geocryological hazard is high not only at southerly latitudes in Canada, but also in northern Nunavut, with low to moderate risk on the mainland. There is high potential for coastal erosion in central and eastern arctic Canada (e.g., Arctic Archipelago), Europe and eastern Siberia, increasing the vulnerability of transportation and pipeline corridors to adverse effects in these locations. Risk of settlement in permafrost soils is less pronounced in mountainous terrain. Potential subsidence in thawing soils of variable ice content, and under various warming scenarios to 2100, are examined by Instanes *et al.* (2005). Generally, the potential for subsidence is lower in soils that are currently colder and of lower ice content (Parmuzin and Chepurnov, 2001).

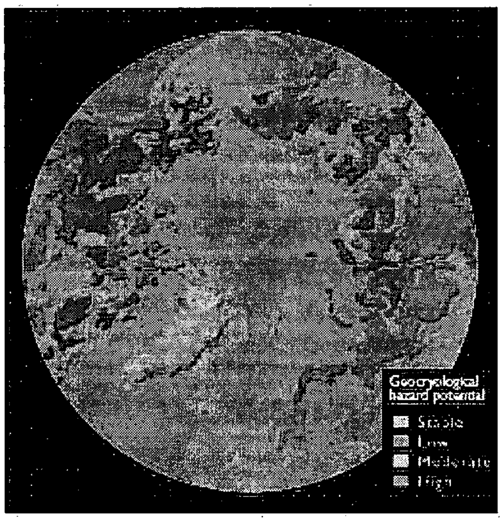


Figure 21. Potential for thaw-induced settlement and other geocryological hazards in engineered structures to 2050 (Instanes *et al.*, 2005, Figure 16.19). These ratings are based upon the HadCM scenario, predicted change in active layer thickness, and ground ice content. Ratings for Nunavut are high in coastal locations and in the Archipelago, and low to moderate inland.

Damages to infrastructure with warming permafrost and increasing thickness of the active layer will affect safety and costs (repair, upgrade, replacement), and will have implications for transportation and environmental quality. Stability of open pit mines in steep permafrost slopes will likely decline as permafrost warms. As mean annual temperatures increase progressively over the 21st century and the active layer thickens, thaw-settlement and creep of piles and footings will increase (Instanes *et al.*, 2005). For example, Nixon (1990, 1994) simulated these effects over a 25 year, 0.1°Cy^{-1} warming, and found a doubling in thaw depth and an increase in creep settlement (e.g., warming to 3 m depth along a pile). Even slight increases in mean annual temperature (i.e., fraction of a degree) can, therefore, result in failure of foundations, particularly those with shorter piles, thereby decreasing the stability, safety and life of structures (Khrustalev and Pustovoit, 1993). Future permafrost warming will also likely cause deformations or breaks in transportation routes (rail lines, pipelines, runways). Already, within the Russian North, there has been dramatic thermokarst activity resulting in collapse of, and damage to, buildings, runways and power stations (Nelson *et al.*, 2002).

Measures to adequately deal with permafrost-induced damage to infrastructure will weigh heavily on the economics of some regions. These measures include: assessment of state or risk of structures in various locations; mitigation planning (from demolition for safety reasons to water quality protection measures); modification of structure design (e.g., redesign of pipelines to account for frost heave and thaw settlement, Nixon, 1994; Nixon *et al.*, 1990a); adaptation of maintenance and construction approaches (Instanes *et al.*, 2005). Costs will further increase due to poor design, construction and maintenance of current structures, which has been shown to account for 45% of deformations in buildings in the past (Kronik, 2001). Khrustalev and Pustovoit (1993) state that construction of new structures as old ones age and become abandoned will be an inadequate response to the expected exponential rise in need for new structures in the future (i.e., 5% per decade over 1980-1990, and 108% per decade in 2030 to 2040). Application of predictive maps of geocryological hazards will help to identify those areas that are at risk, and potentially reduce future costs by allowing pre-emptive measures to be taken (e.g., Anisimov and Belolutskaia, 2002; Nelson *et al.*, 2001; Instanes, 2000).

The effects of permafrost change on infrastructure will not occur in isolation; changes to arctic hydrology will be important as well. Though potential damages to infrastructure due to changes in permafrost may be predicted using coupled climate-soil models and geocryological hazard maps, interactions and feedbacks between temperature, precipitation, soil and permafrost, vegetation, ice and discharge, extreme events and local conditions confound these projections. As changes to precipitation and the ice-water balance of the active layer progress, changes to runoff, including distribution and magnitude of discharge and flooding regimes, will vary regionally and locally, as will effects on infrastructure. With increased frequency and magnitude of storms and increased river discharges, infrastructure designed for historic extremes may be damaged, and new criteria for location and design may need to be developed (Instanes *et al.*, 2005). As ice covers on lakes and rivers thin, loadings on bridge piers and offshore structures will decline, and ice road use and design will be more limited. Heavier, more frequent precipitation and soil thaw will increase the occurrence of slope failures, development of thermokarst sink holes, slides, mudflows, and wet snow and slush avalanches, which may damage infrastructure as well as pose a risk to human life (Sanderson *et al.*, 1996; Hestnes, 1994). For example, catastrophic mudflows in Sima valley (Norwegian West Coast, 1893 and 1937) and Tyznuays (Caucasus, 1977 and 1992) were generated by heavy rain storms and intense snow melt (Seinova, 1991; 1998). Extreme events such as mudflows will, however, be short-lived and should stabilize over time.

Infrastructure in coastal locations will be affected not only by changes to permafrost and freshwater hydrology, but by changes to coastal processes as well. Warming and storminess are projected to increase most dramatically in coastal locations of the Arctic (Källén *et al.*, 2005). Coastal erosion associated with degrading permafrost (e.g., thaw subsidence in low-lying, ice-rich permafrost) will be exacerbated by greater wave energy associated with extended ice-free period and more intense and frequent storms, as well as a more dynamic ice cover. Sea level rise due to thermodynamic expansion of marine waters and melt of glaciers and ice caps will also accelerate erosion and increase flooding, leading to loss of coastal land in some locations, and increasing salinity of coastal freshwaters.

VIII. SUMMARY

Annual air temperatures in the Arctic are projected to rise by between 2-5°C over landmasses (3-7° for the winter season) and by between 5-10°C over large water bodies (6-10° over winter) by 2071-2090. Average annual precipitation is projected to increase by approximately 12.3% for the whole Arctic, with greater increases over large water bodies than over landmasses, and with net annual water availability (P-E) increasing by 15.5% by 2071-2090. Coastal areas in the Arctic will, therefore, likely be warmer and wetter than areas further inland under future climate conditions. These changes will be regionally and seasonally variable, with the greatest increases in air temperature and precipitation associated with the cold season.

The climate of Nunavut is expected to be warmer and wetter under the projected future climate regime. Air temperatures of terrestrial Nunavut are projected to increase on average by 4.7°C for the cold season, and by 2.6° for the warm season by 2071-2090, with precipitation increasing by 15.1 and 21.4 mm, respectively. These changes over the cold and warm season will be accompanied by increases in the number of degree days, hence a lengthening of the warm season, and increased potential evaporation. Much like other areas of the Arctic, the greatest degree of warming and the most pronounced increases in precipitation are projected to occur over the fall, winter and spring seasons, as well as over and adjacent to large water bodies and coastal areas. The most pronounced changes are likely, therefore, to occur at higher latitudes in association with the projected warming over the Arctic Ocean. Changes in fall temperatures are projected to increase rapidly at higher latitudes, with extreme increases in temperatures over the northern Archipelago into early winter (near 10°C in November) and less pronounced warming further inland (3-6°C). As the winter season progresses, this pattern is expected to shift latitudinally, with temperature increases becoming less extreme and more uniformly distributed. Over the warm season, air temperatures are expected to increase more greatly over central Nunavut (near 3°C) and to a lesser degree over the northern Archipelago (near 0°C). Increases in winter precipitation are expected to be heterogeneous in distribution (10 to 15%), though generally greater in areas adjacent to Hudson Bay and Foxe Basin, and in the northernmost Archipelago (near >25%).

Sea ice extent is projected to decline through the 21st century. The greatest decline in sea ice coverage is projected to occur in the fall, increasing the length of the open-water season. Winter ice extent will likely not change considerably, though the ice edge will retreat, reducing the amount of land-fast ice at high latitudes. Multi-year ice is projected to decline in coverage. Reduced distribution of ice (annual and perennial) within Nunavut will likely increase precipitation and cloudiness, coastal storms, erosion, navigability of the Northwest Passage, and offshore activities.

Glaciers and ice masses are projected to decline in mass with future warming, contributing to sea level rise. Glaciers at lower latitudes, which are typically more sensitive to changes in temperature and precipitation, will experience the greatest mass loss. Ice masses of the Canadian Arctic, which are generally less sensitive to mass loss than those in other parts of the Arctic, will potentially contribute 0.5 cm to sea level by 2100, which is a fraction of that projected for Alaska and the Greenland Ice Sheet. Based on past records, this loss of mass will likely be due to rising temperatures and melt rather than to a decline in accumulation. Glacial melt water flows will, therefore, be expected to rise, affecting the hydrology of lakes and rivers and, via impacts on sea-level, affecting coastal communities.

The extent and duration of snow cover for much of the Arctic is projected to decline in the future despite expected increases in precipitation. The greatest reductions in snow cover will likely be at low latitudes, and will occur in the spring in response to rising temperatures and albedo-feedback effects as the ground becomes free of snow earlier. Similar effects will occur in autumn in locations where snow cover is less abundant. Though the duration and extent of snow cover will decline under future climate scenarios, snow water equivalence may increase as atmospheric moisture at high latitudes increases, particularly in coastal locations where open-water may develop in perennially ice-covered marine waters.

As climate warming progresses over the 21st century, the areal extent of the continuous permafrost zone will contract by near 40%, with the sporadic and discontinuous zones shifting northward, and the depth of the active layer increasing by between 30 and 50 %. Increases in active layer depth are projected to be greatest at higher latitudes in Nunavut. Continuous permafrost in the territory is projected to remain relatively intact in distribution over the 21st century, with active layer depth increasing by greater than 50% by 2050. At lower latitudes within the territory there will be a slight northward shift in the continuous permafrost zone and slightly less pronounced deepening of the active layer. Effects of permafrost thaw over the last century have included increased erosion and landslides, damage to infrastructure, degraded water quality, as well as major shifts in vegetation, which have implications for feedbacks to the climate system through alteration of greenhouse gas exchange. Building collapse and damage, and damage to runways and power stations, are already an issue in the Russian North, where thermokarst activity has been aggravated by recent warming (Nelson *et al.*, 2002).

River and lake ice of Nunavut may be expected to melt earlier in spring and freeze later in the fall based on projected changes in air temperatures. Ice thicknesses will generally be thinner than those of present-day. Ice composition may also be expected to change. Impacts include a likely reduction in water levels, decreased severity of break-up and flooding, and reduced quality and distribution of ice, hence deterioration of ice roads and potentially increased opportunity for river-based transport.

Water quality in Nunavut may be expected to decline in the future as contaminant burdens and loadings of sediment, nutrients and organic carbon increase. Contaminant burdens in melt water runoff and flood waters will increase with increased atmospheric deposition, melt of perennial cryospheric stores, contamination of water supplies due to permafrost- and floodwater-induced damage to water distribution and sanitation infrastructure, and increased human activity under a more moderate climate regime. Runoff loadings and erosion-derived sediment and organic carbon may not only degrade water quality due to contaminant contributions, but also due to increased turbidity of lake and pond waters. This, along with temperature- and nutrient-induced increases in productivity (and potential eutrophication), will increase the need for modification of existing, and development of new, treatment systems for water supplies in Nunavut. Treatment for saltwater contamination of freshwater supplies may be of particular concern in coastal areas.

Deterioration of permafrost, shifts in hydrology and increased frequency and magnitude of extreme events will threaten the integrity of infrastructure. As permafrost warms and the active layer thickens in Nunavut, soils will lost their strength, thereby increasing creep, thaw settlement and frost-heave. Structure foundations will be at risk of damage, particularly in locations that are at high risk of geocryological hazard (i.e., thaw-induced settlement), including northern Nunavut. Geocryological hazard projections, as well as simulations of future freeze and thaw indices, aid forecasts of potential change in permafrost and risk of damage, which in turn may be incorporated into infrastructure design. Damage to existing structures with shifts in permafrost, and damages due to increasingly frequent and heavy storms and associated e.g., mudflows and coastal erosion, will involve costly repairs and may result in loss of life.

IX. ACKNOWLEDGMENTS

The author would like to thank:

Terry Prowse (National Water Research Institute, Victoria, BC) and Barrie Bonsal (National Water Research Institute, Saskatoon, SK) for reviewing this report;
Bill Chapman (Department of Atmospheric Sciences, University of Illinois) for providing GCM data for Nunavut.

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