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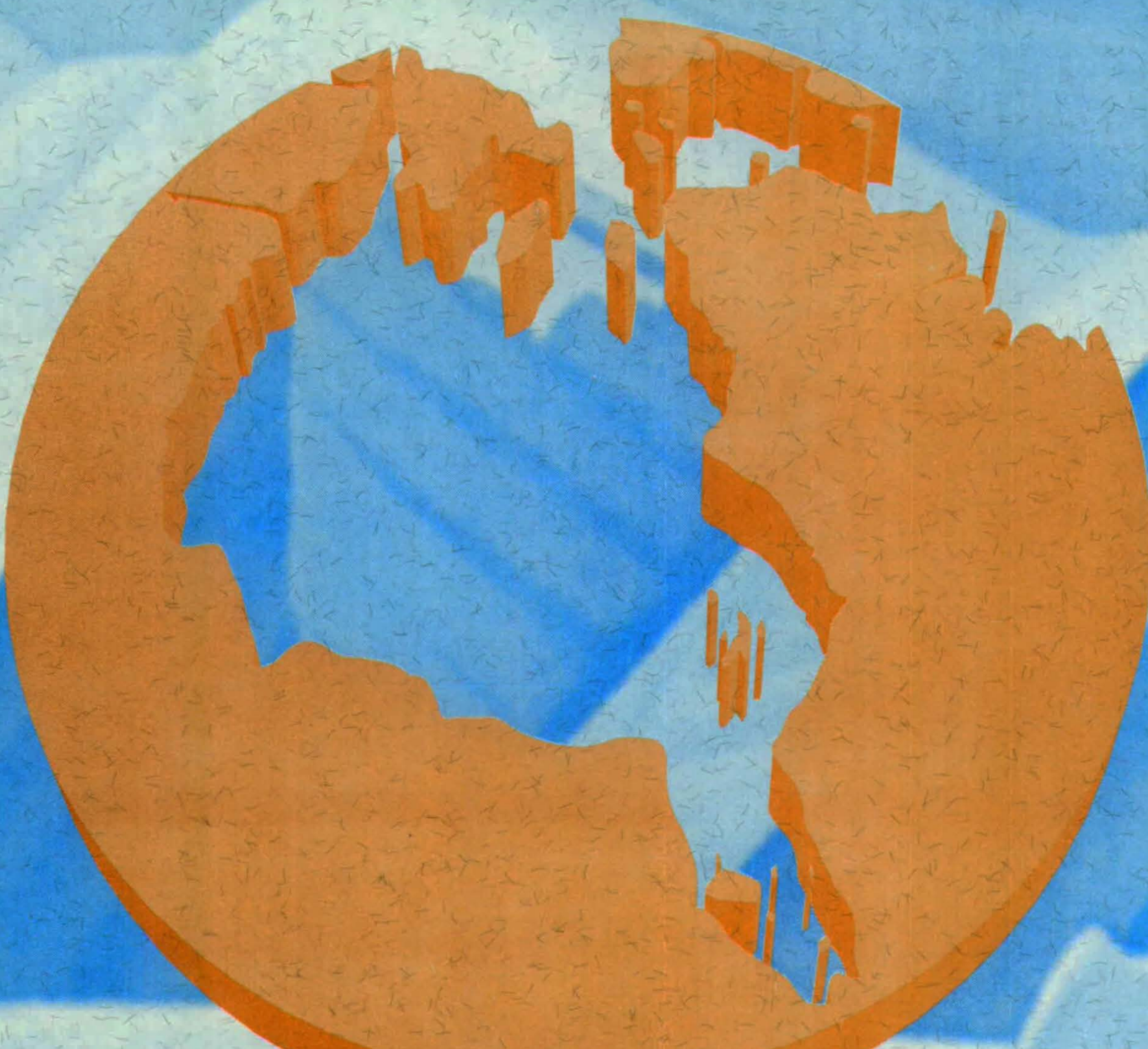

HUDSON BAY PROGRAMME SUR LA BAIE D'HUDSON

CLIMATE VARIABILITY, CLIMATIC CHANGE, AND IMPLICATIONS FOR THE FUTURE OF THE HUDSON BAY BIOREGION

Stewart J. Cohen, Tom A. Agnew,
Angus Headley, Paul Y.T. Louie,
Jenny Reycraft, and Walter Skinner

Environmental Adaptation Research Branch
Atmospheric Environment Service
Environment Canada
Downsview, Ontario M3H 5T4

The Hudson Bay Programme

1994

Canadian Cataloguing in Publication Data

Climate variability, climate change and implications for the future of the Hudson Bay bioregion

Includes bibliographical references.
ISBN 0-919996-56-6

1. Hudson Bay Region--Climate. 2. Climatic changes--Hudson Bay Region. I. Cohen, Stewart Jay. II. Hudson Bay Programme.

QC985.5.H83C54 1994 551.69163'27 C94-900255-0

The Hudson Bay Programme 1994
Cover design, Graphix Design
Printed by the Canadian Arctic Resources Committee

This report was commissioned by the Hudson Bay Programme, however the views expressed herein are those of the author and do not necessarily represent those of the partners in the Hudson Bay Programme.

The Hudson Bay Programme acknowledges the financial support of: The Richard and Jean Ivey Fund, The Harold Crabtree Foundation, Walter and Duncan Gordon Charitable Foundation, The McLean Foundation, George Cedric Metcalf Charitable Foundation, Helen McCrea Peacock Foundation, Murphy Foundation Incorporated, John D. and Catherine T. MacArthur Foundation, The Molson Family Foundation, The EJLB Foundation, Government of the Northwest Territories, Indian and Northern Affairs Canada, Environment Canada, Environmental Innovation Program, Manitoba Hydro, Grand Council of the Crees (of Quebec), Mushkegowuk Tribal Council, Ontario Hydro, Hydro-Québec.

SUMMARY

The climate of the Hudson Bay Bioregion is reviewed. Patterns of air temperature, precipitation, ice cover, soil temperature, and streamflow are described for the instrumental record, and presented in graphs and maps. Recent climatic trends are noted. Scenarios based on climate model projections of greenhouse gas-induced climatic change are discussed, as are their potential implications. There are uncertainties associated with projections of climatic change impacts, so the reader is advised to consider this as a scenario, not a forecast. This review has also been hampered by a sparse monitoring network and incomplete or missing data records.

Highlights of this review are as follows:

Current Regional Climate:

The region's cold climate is largely a result of its high latitude continental location, far removed from the Pacific Ocean and other sources of milder air. The Bay plays only a secondary role in determining the position of the upper air trough which controls atmospheric circulation in this region.

The cold climate has led to the development of discontinuous and continuous permafrost everywhere except south of James Bay. This, along with high snowmelt and the presence of ice in lakes and rivers, influences the regional hydrologic regime, which encourages large spring peak flows and the development of wetlands in lowland areas.

Sea ice is the largest component of the freshwater budget of Hudson Bay. It also provides the physical controls on biological productivity, distribution, and interaction for all levels in the food web.

Climate Trends:

This region is located in a transition zone between an area of anomalously warm conditions (northwestern North America) and an area of colder than normal conditions (northern Atlantic Ocean). Certain regional trends are evident, particularly the warming in spring; however, these trends are of a lesser magnitude than current variability. There is no clear regional precipitation trend, but data analysis is affected by snowfall undercatch and changes in instrumentation. Some lakes in the south and southwest are exhibiting earlier breakup of ice, and an earlier spring peak is being observed at the Missinaibi River. Recent cooling has occurred in the Baffin Island region, but this zone is too far east to be noticeable in the Hudson Bay region.

Climate Scenarios:

The Canadian Climate Centre's General Circulation Model (CCC-GCM2) simulation of CO_2-induced climatic change shows higher temperatures in all seasons, accompanied by the disappearance of sea ice from the Bay. There are uncertainties associated with current GCMs, particularly the lack of a fully circulating interacting ocean and poor treatment of clouds.

Implications of Scenarios of Climatic Change:

Climatic changes could have major implications for ecosystems of the Bay, but few impact studies have been completed to date. What can be said is that there are climate-sensitive areas that could be affected by relatively small increases in temperature, particularly i) permafrost warmer than -2°C, ii) coastal areas up to 10 km inland, iii) wildlife dependent on sea ice and wetlands, iv) infrastructure built in the discontinuous permafrost zone, and v) communities dependent on wildlife. It should also be noted that scenarios of climate warming, hydroelectric development, and the GRAND Canal would probably lead to very different impacts on the bioregion. For example, warming might lead to a reduction in sea ice, but the GRAND Canal's freshwater impoundment, with its reductions in salinity, could have the opposite effect.

Research Needs:

1) There should be increased monitoring of snow cover, streamflow in unregulated rivers, freshwater ice, the Bay's freshwater budget (including snow cover and snow melt over the sea ice) and ground temperatures, so that significant changes in permafrost, hydrology, and salinity could be detected. This would complement ongoing efforts at monitoring temperature, sea ice, and other parameters. These efforts should be protected and maintained over the long term.

2) An integrated regional assessment of scenarios of climatic changes and water resources development should be carried out. This could focus on climate-sensitive areas and their various stakeholders (governments, communities, resource industries, ecosystem maintenance, transportation, tourism).

ACKNOWLEDGEMENTS

This report was a group effort, coordinated by Stewart J. Cohen, who also served as editor. Lead authors for each chapter of this report are listed below:

1. Tom A. Agnew
   - 1.4.2 Walter Skinner
   - 1.5-1.5.2 Angus Headley
   - 1.6-1.7 Paul Y.T. Louie

2. Tom A. Agnew and Walter Skinner
   - 2.6 Paul Y.T. Louie

3. Walter Skinner

4. Stewart J. Cohen
   - 4.2 Angus Headley

5. All

Computer-generated figures were produced for each chapter as follows:

1. Tom A. Agnew
   - 1.5 Angus Headley
   - 1.6 Paul Y.T. Louie

2. Walter Skinner and Jenny Reycraft
   - 2.6 Paul Y.T. Louie

3. Walter Skinner and Jenny Reycraft

Appendix 1 Tom A. Agnew and Angus Headley

Appendix 2 Walter Skinner and Jenny Reycraft

Appendix 3 Walter Skinner and Jenny Reycraft

In addition, Krystyna Czaja assisted in figure reproduction and layout.

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1.1 Introduction

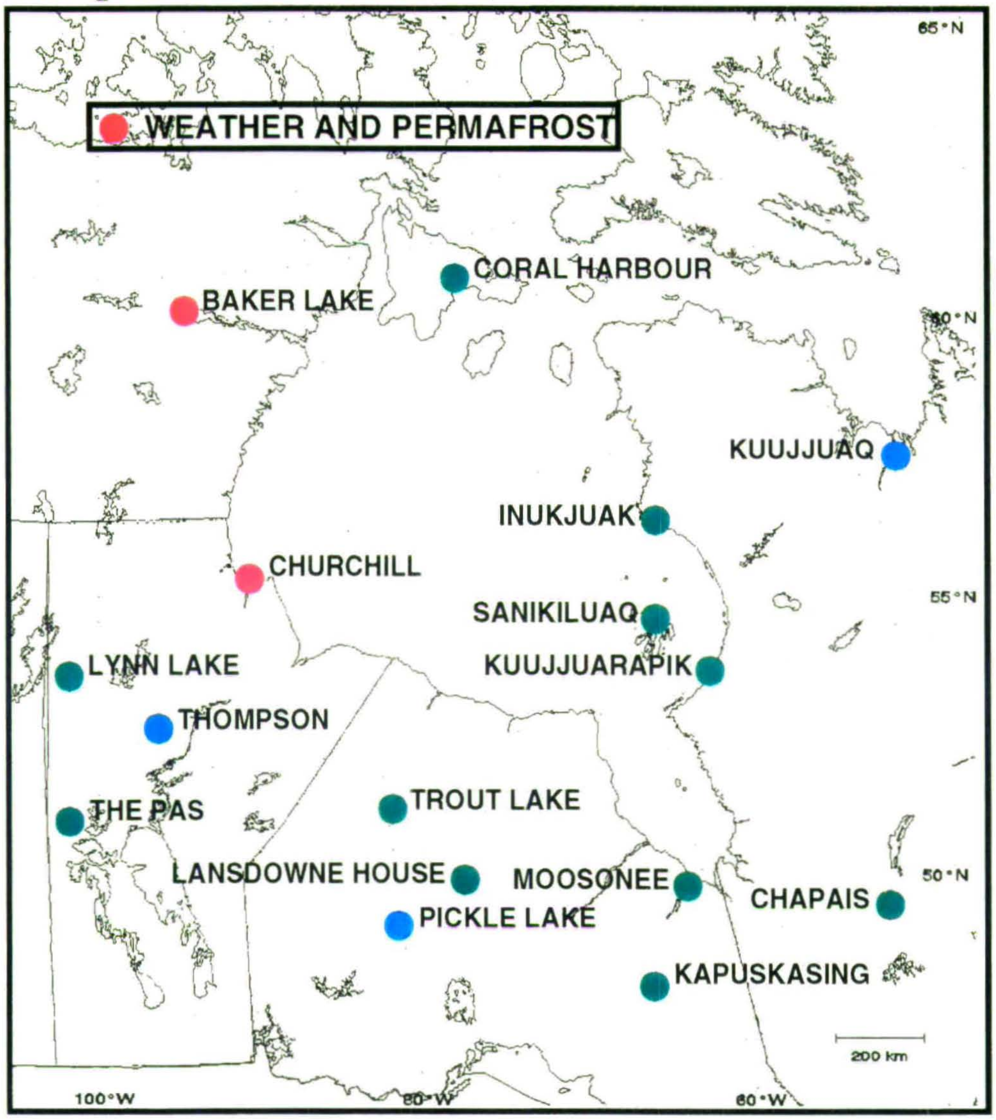
The Hudson Bay Bioregion is located within the sub-arctic and arctic regions of Canada. The climates of these regions are characterized by long cold winters and relatively short growing seasons. A number of landscape features reflect this, particularly the presence of discontinuous and continuous permafrost, freshwater and marine ice, and the arctic tree line, which represents the northern limit of the Boreal forest.

Hudson Bay and James Bay combined is the largest inland sea in the world. Surrounded by the North American continent on the west, south, and east, exchanges of water with the Atlantic occur through Hudson Strait. A negligible exchange with the Arctic Ocean occurs through the Foxe Basin and Roes Welcome Sound. As a result, it is to a large extent a self-contained body of saltwater. Its climate is influenced by the North American continent and as a result is known for its extremes of cold in winter. It in turn has considerable influence on the regional climate of the area. In this summary, Hudson and James Bay will often be referred to as Hudson Bay or simply the Bay.

Early records of weather observations were taken by fur traders at a number of posts around the Bay during the eighteenth century. However, most of these observations were intermittent. It was not until Churchill was opened in 1931 as a terminus for a northern sea route to transport grain that a comprehensive record of weather observations began at locations around the Bay. These weather observing stations were required to provide weather and sea state forecasts to support shipping and other activities of the area. Over the years, this information has been used to summarize the climate of Hudson Bay by a number of authors (Danielson 1969; Maxwell 1986). In addition sea ice climatologies have been produced by Markham (1986, 1987) based on ice reconnaissance over the years and the oceanography, including the heat and freshwater budget of the Bay, has been summarized by Prinsenberg (1984, 1986a, 1986b, and 1988) using oceanographic research information. This report will to a large extent summarize these studies and in addition use more recent data to discuss climate variability and trends.

The station data used in this summary are long term Atmospheric Environment Service (AES) weather observing stations. Although 14 land stations surrounding the Bay are used, for ease of discussion, graphs of a representative station are shown in the body of the text. The station locations are shown in *Figure 1.1.1*. Station precipitation amounts are uncorrected for undercatchment problems associated with aerodynamic flow around

Figure 1.1.1. Base map of the Hudson Bay area with weather observing stations (●) and permafrost sites (●).



the measurement container and container wetting losses (Metcalfe and Goodison 1993).

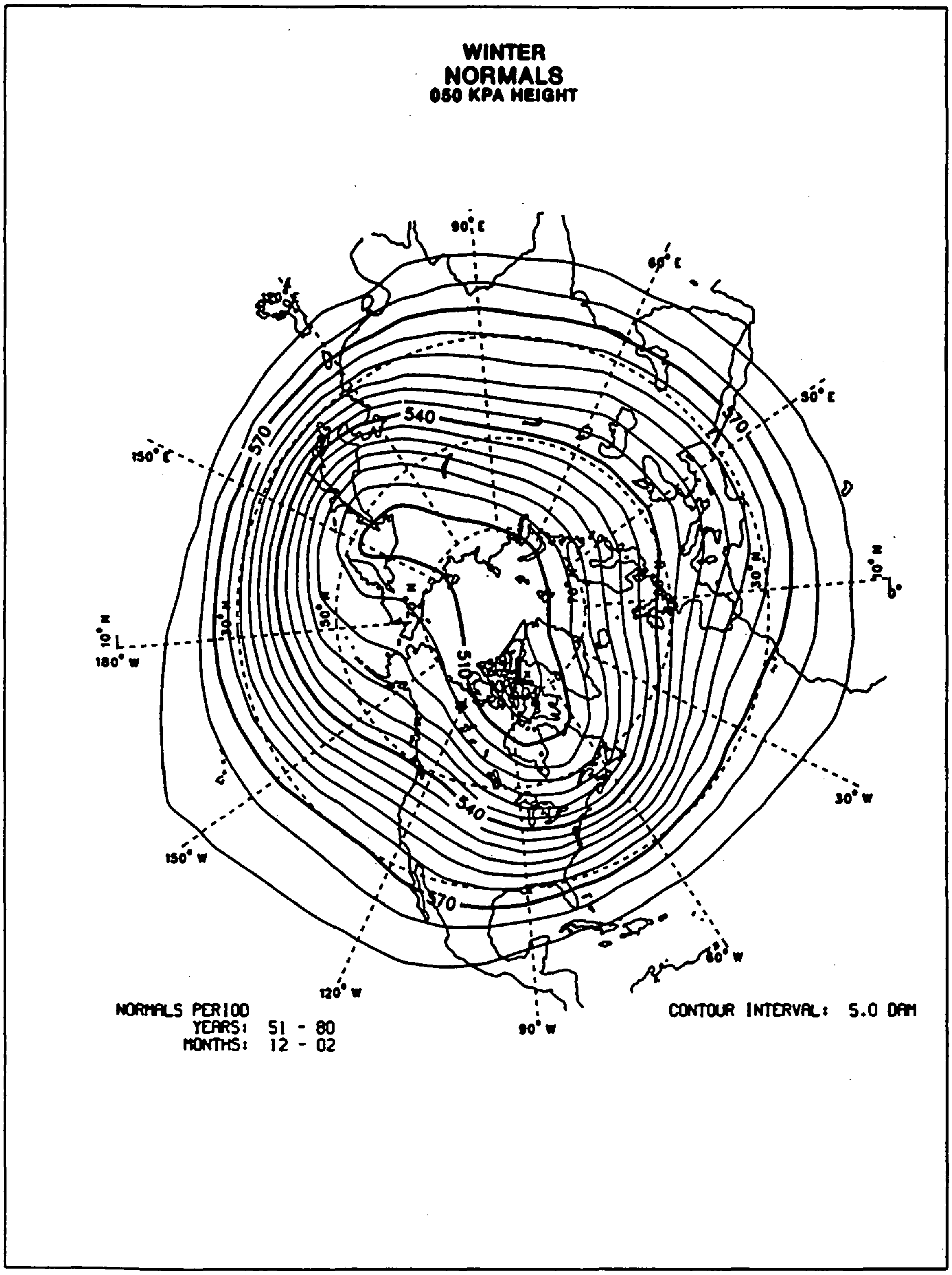
Marine summary information over the Bay is obtained from the Comprehensive Oceanographic and Atmospheric dataset (COADS, Woodruff et al. 1987) and the Canadian Cooperating Ship data. Except for specific oceanographic research data these data are the only information available over the Bay. These data are used to summarize the marine climate; however, they have the following limitations: 1) the observations are few in numbers and are limited to the open water season, 2) they are not evenly distributed over the Bay, and 3) the observations are done visually by untrained observers. The sea ice data are obtained from digitized composite sea ice charts prepared by the Ice Branch of AES (Markham, 1986 and 1987) and from the US NOAA/NAVY Joint Ice Centre digitized hemispheric sea ice charts (Knight 1984). These data cover only the period 1972 to 1988. Summaries for various parameters using these datasets are produced by two software packages developed by AES called Marine Analysis Statistics (MAST) and Climate Research in Ice Statistics Package (CRISP) described by Saulesleja et al. (1985) and Agnew et al. (1987).

The motivation for this report comes from the recent concern over cumulative effects of existing and proposed large scale hydroelectric development in the area in conjunction with concerns over climate change. It has been produced at the request of the Canadian Arctic Resources Committee (CARC).

1.2 Climate Controls

1.2.1 Large Scale Circulation and Weather Systems

The upper-air circulation controls the type and frequency of air masses and weather systems which move over the Hudson Bay area. The winter and summer mean 50 kPa height contour for the Northern Hemisphere (*Figures 1.2.1 and 1.2.2*) shows a trough near or slightly to the east of Hudson Bay in both summer and winter. The intensity of the trough is considerably stronger in winter with the deepest part of the polar vortex over the Canadian Arctic Islands. The mean position of the trough is a result of the physical geography of the Northern Hemisphere (i.e. the distribution of the oceans and continents around the hemisphere and the western mountain ranges of North America) and the wave nature of the large scale atmospheric flow. From the perspective of air mass movement the mountain ranges of western North America restrict the movement of warm moist air from the Pacific in favour of arctic air. In terms of atmospheric dynamics, the western mountain ranges of North America set up large scale planetary (Rossby) waves with Hudson Bay approximately one half of a Rossby wavelength downstream. The



1.2.1 Mean winter upper air (50 KPa) circulation for the northern hemisphere.

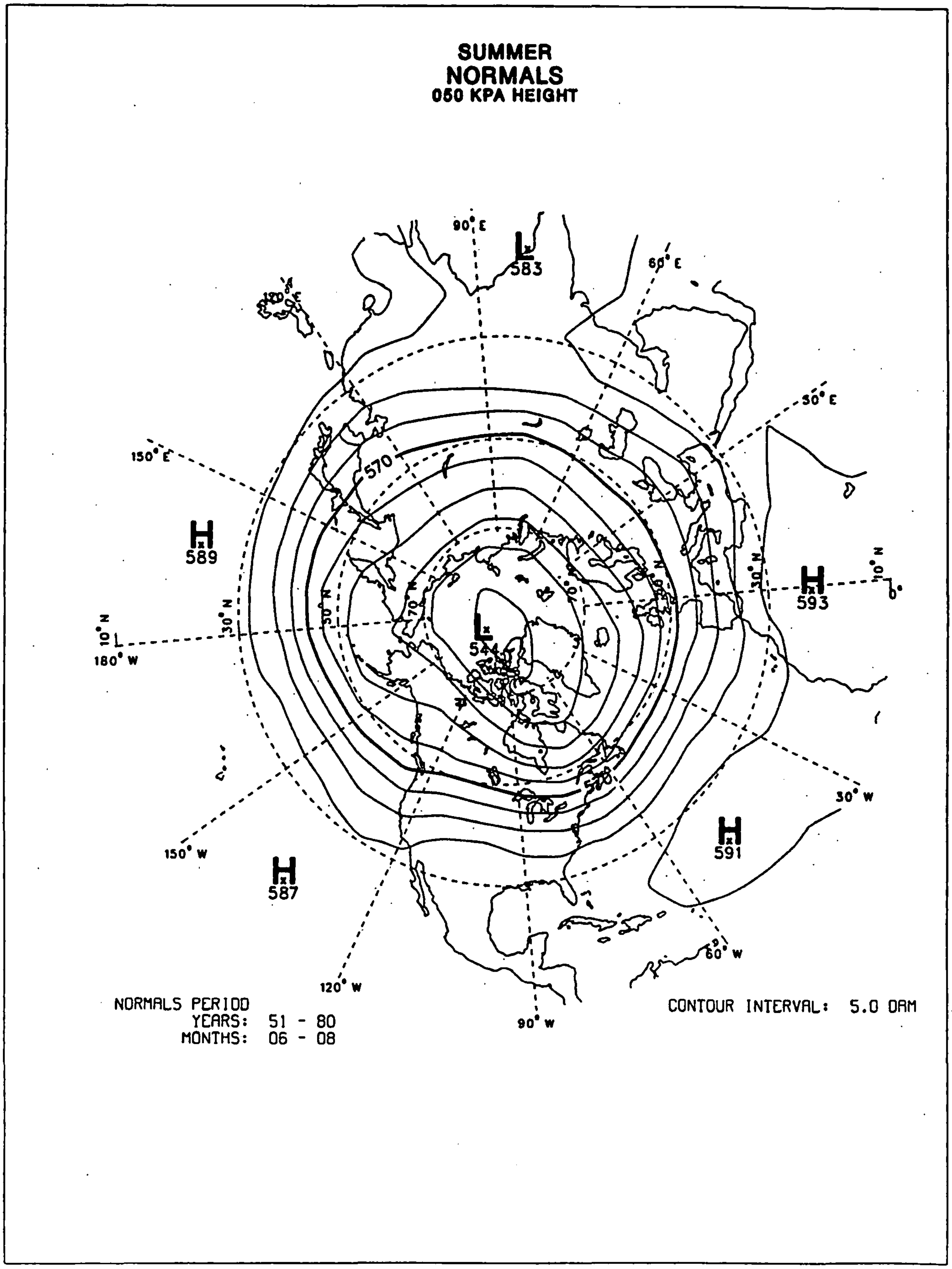


Figure 1.2.2 Mean summer upper air (50 KPa) circulation for the northern hemisphere.

thermal effect of the Bay itself may also be a factor in the position of the trough although this is felt to be of secondary importance (Danard 1980).

As a result of the trough position, winter storms tend to remain south of the Bay affecting the Bay only when they move into Davis Strait on the east side of the trough. During summer the trough weakens and moves northwards and the mean flow tends to be more often south westerly and southerly. With this, storms now often move directly across the Bay bringing relatively mild moist southerly air over the area. This warmer air, however, is considerably modified by the cold waters of the Bay.

The variation in mean position and intensity of the trough has a large influence on the year to year variability in climate of the area. A more intense trough will bring more frequent cold northerly air, as will a shift in the mean position of the trough east of the area. Weakening and/or a westward shift will bring more frequent southerly warmer air and increased storms. This will also affect the timing of ice freezeup and breakup. These effects are discussed in more detail in the section on climate variability.

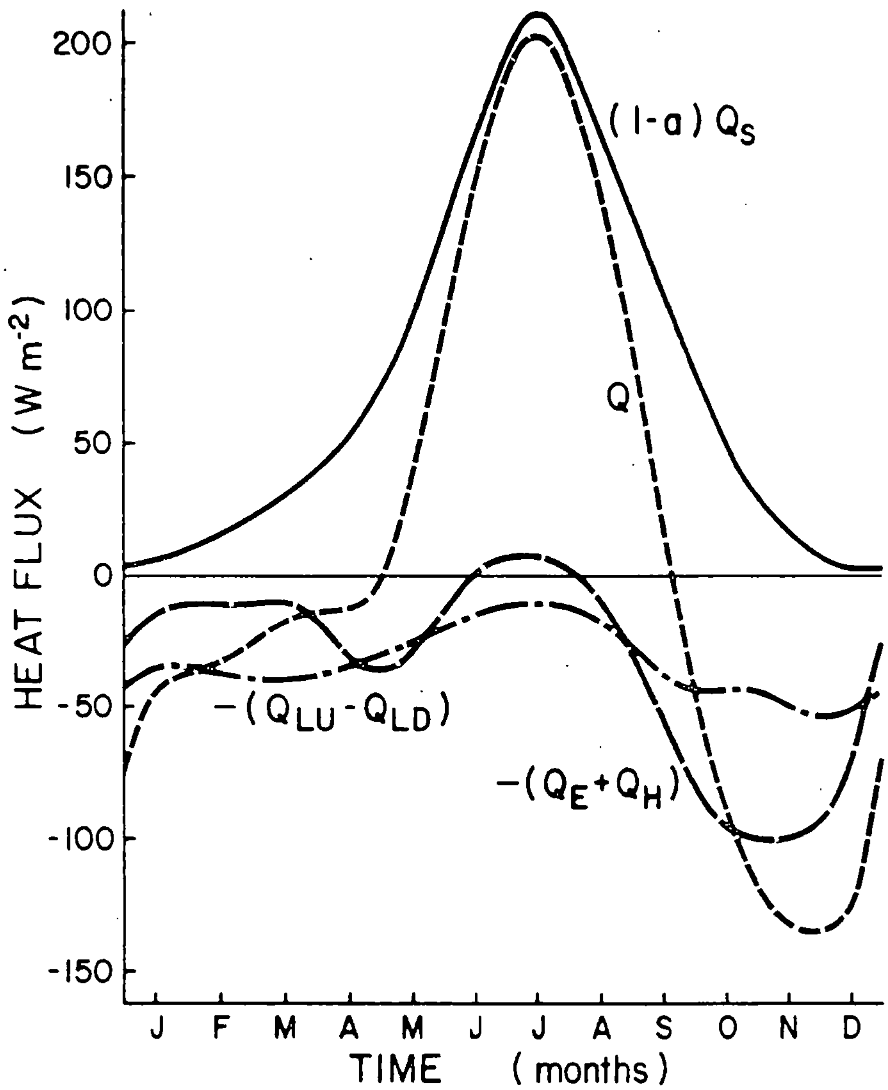
1.2.2 Radiation Regime and Surface Energy Balance

The surface energy balance for Hudson Bay (Figure 1.2.3) is given by Prinsenberg (1986a) as summarized by Danielson (1969). The net heat flux (i.e., the energy available for melting snow and ice surfaces or heating land and water surfaces) is given by

$$Q = \underbrace{(1-a)Q_s + Q_{ld} - Q_{lu}}_{\text{radiative}} \quad \underbrace{-Q_e - Q_h}_{\text{convective}}$$

where $(1-a)Q_s$ is the net solar radiation, Q_{lu} the upward and Q_{ld} the downward longwave radiation, Q_e the evaporative and Q_h sensible heat fluxes and a is the albedo (reflectivity).

The energy fluxes at the surface are usually divided into radiative, convective, and conductive components. The radiative component is dominated by solar radiation from May to August and provides the water with a mean annual heating rate of 78 W/m^2. In September, solar radiation becomes less important and net longwave and convective heat losses dominate as the surface water temperature reaches a maximum and increased frequency of cold northwesterly flow produces unstable convective activity. By January longwave radiation loss is about 35 W/m^2 and this continues over the rest of the winter. The sea ice growth rate over the winter period depends on the rate of conductive heat loss. The ice usually increases in thickness until around April. At that time increased solar radiation begins to dominate the



1.2.3 Monthly averaged surface energy balance (from Danielson, 1969). The net energy balance Q is made up of the solar radiation (1-a)Q, upward and downward longwave radiation (Q_{lu} - Q_{ld}), and evaporative and sensible heat fluxes Q_e and Q_h, respectively.

surface energy balance again. About half of the solar energy input during summer is used to melt the sea ice cover and the rest is used to heat the waters of the Bay.

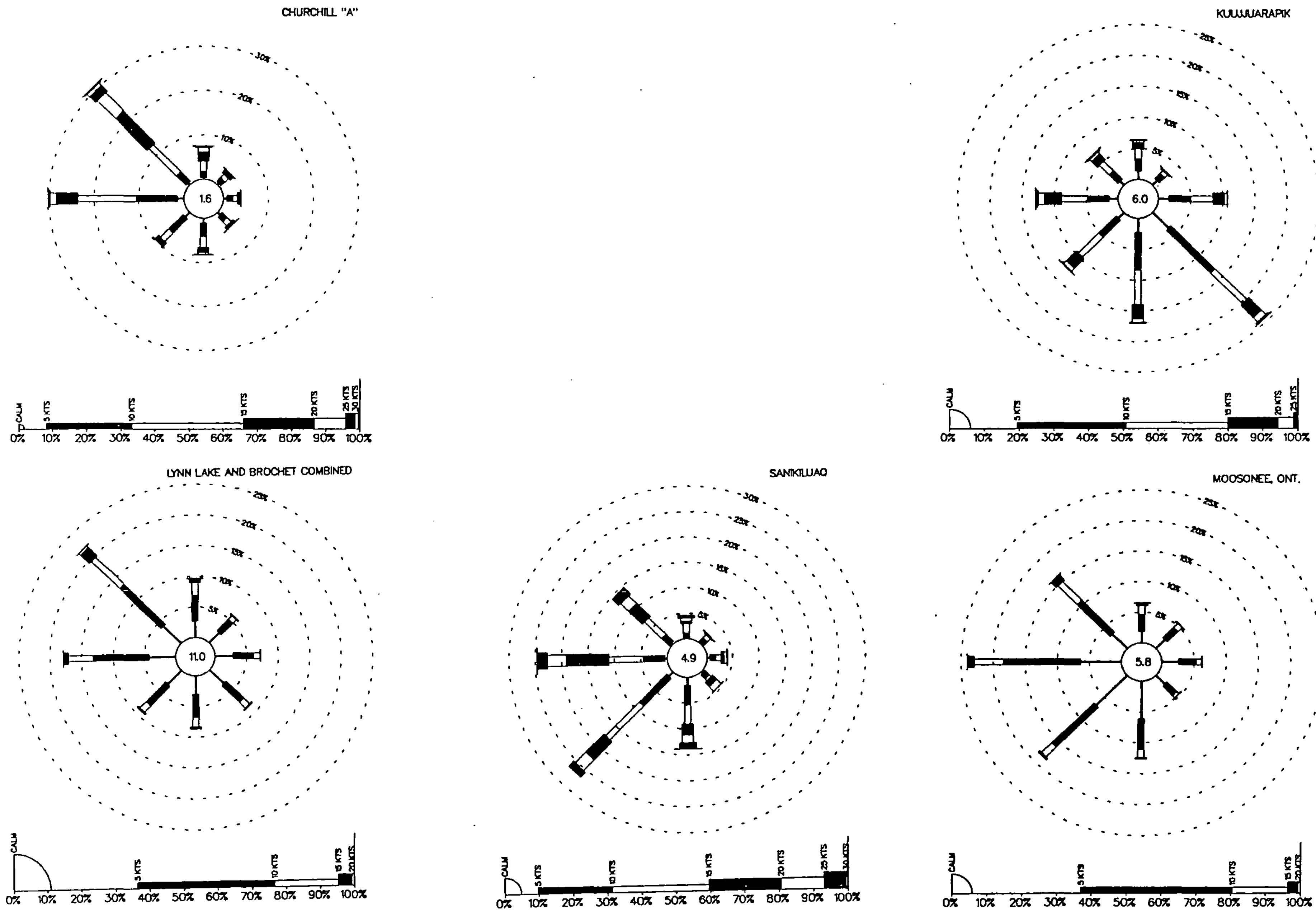
1.2.3 Nature of Immediate and Adjacent Surfaces

The changing nature of the land and water surfaces in the area affects regional climate. Water bodies are either ice covered with little contrast between land and water, or cold open water. During winter, the sea ice cover causes the water to experience the same climate as the snow-covered tundra that surrounds it. This allows outbreaks of cold arctic air to penetrate deep into the North American continent. In summer, the very cold waters limit air temperature in the immediate coastal area to less than 10°C.

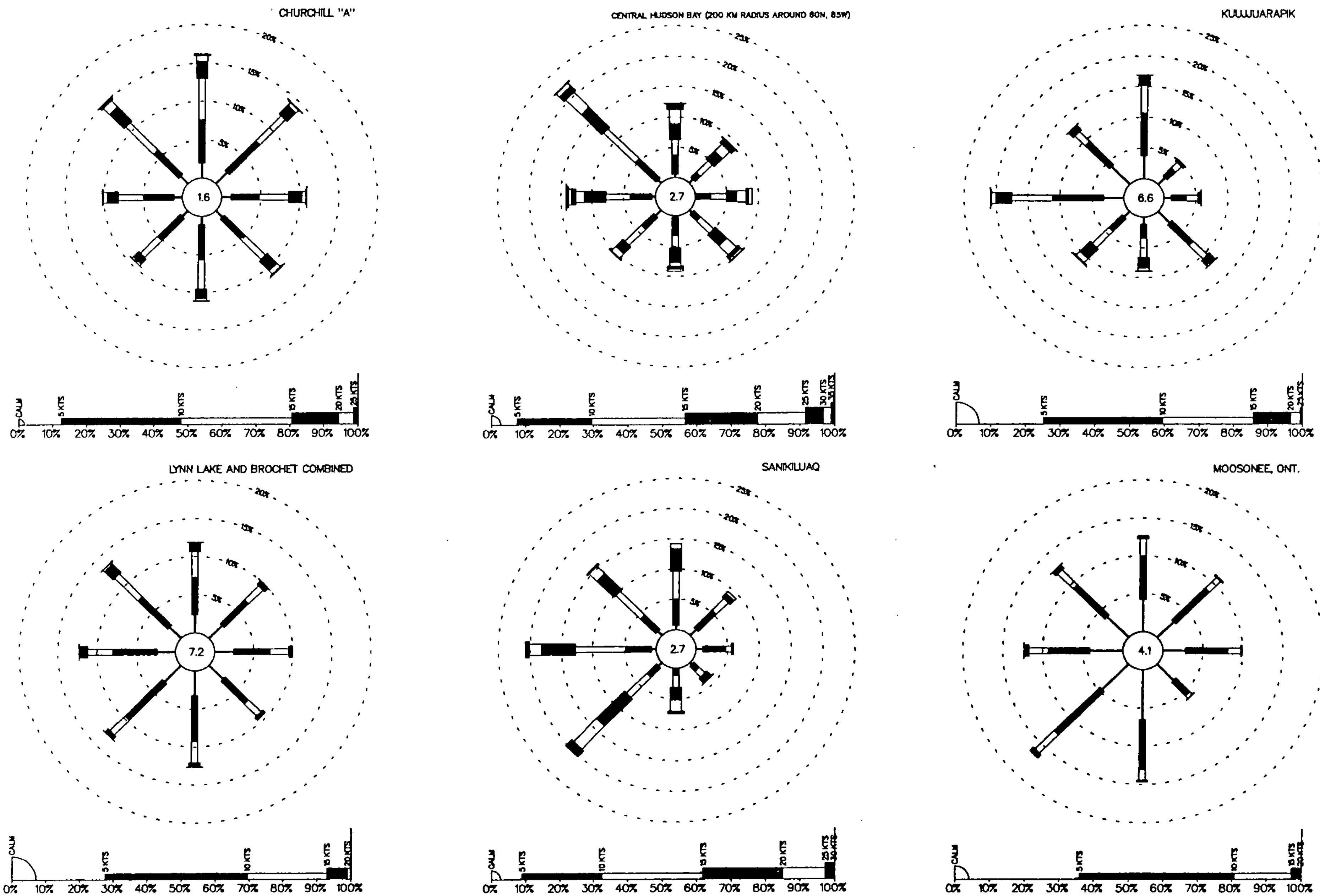
All of the coastal lands around Foxe Basin and much of the area surrounding Hudson Bay are characterized by continuous permafrost. Discontinuous permafrost can be found along the south shores of Hudson Bay and around James Bay (see section 1.5). The presence of permafrost has an important impact on summer climate of the area especially to the west of Hudson Bay which is at least partially responsible for the water-logged state of the terrain. Water from melting snow and spring rains is unable to penetrate the frozen ground but simply collects in puddles and small ponds. The result is a wet swampy surface where evaporation consumes heat energy in the summer which could otherwise be available to increase air temperature. In late summer, water temperatures are at a maximum and air-water temperature differences produce enhanced precipitation on the eastern side of the Bay.

1.3 Regional Climate

The Bay extends over a large area and climate differs between more northerly and southerly parts of the Bay and on the western and eastern side of the Bay. In terms of the wind regimes, northwestern parts of the Bay have predominantly northwesterly flow especially in winter with the eastern side of the Bay predominantly south to southwesterly (see Churchill versus Sanikiluaq and Kuujjuarapik, *Figure 1.3.1*). This is mainly a result of the position of the upper level trough and the strong surface pressure gradients maintained by the persistent surface high pressure over the Keewatin District and the low pressure over the Davis Strait/Labrador Sea. During summer, winds are more variable as the upper flow weakens (*Figure 1.3.2*). Precipitation rates also differ considerably over the region, with total precipitation in James Bay near double that of Hudson Bay, and precipitation on the western side of the Bay about 40% higher than on the eastern side in summer (*Figure 1.3.3*). These



1.3.1 Winter wind roses for selected stations around Hudson Bay.



1.3.2 Summer wind roses for selected stations around Hudson Bay.

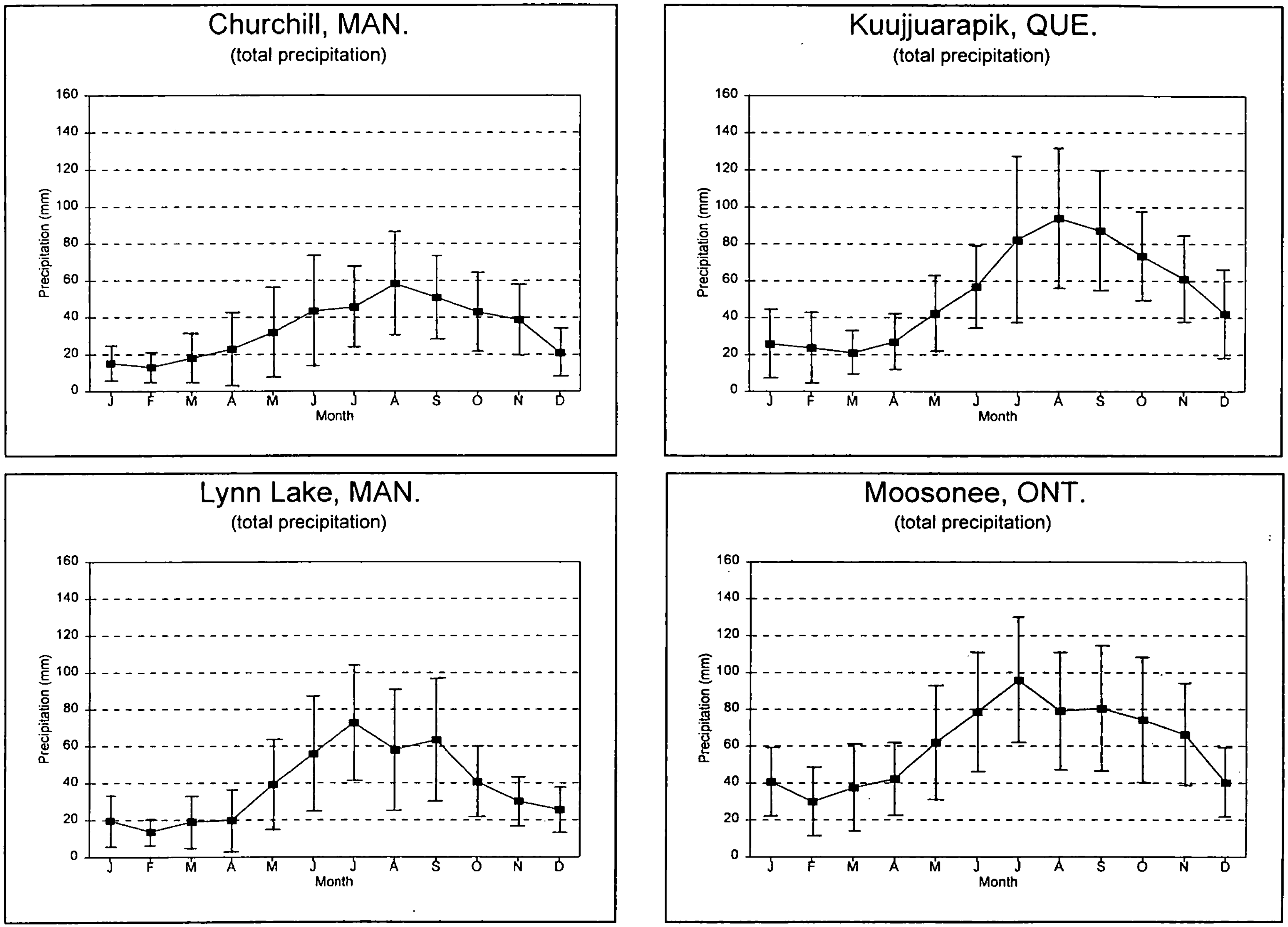


Figure 1.3.3 Monthly precipitation for locations around the Hudson Bay region.

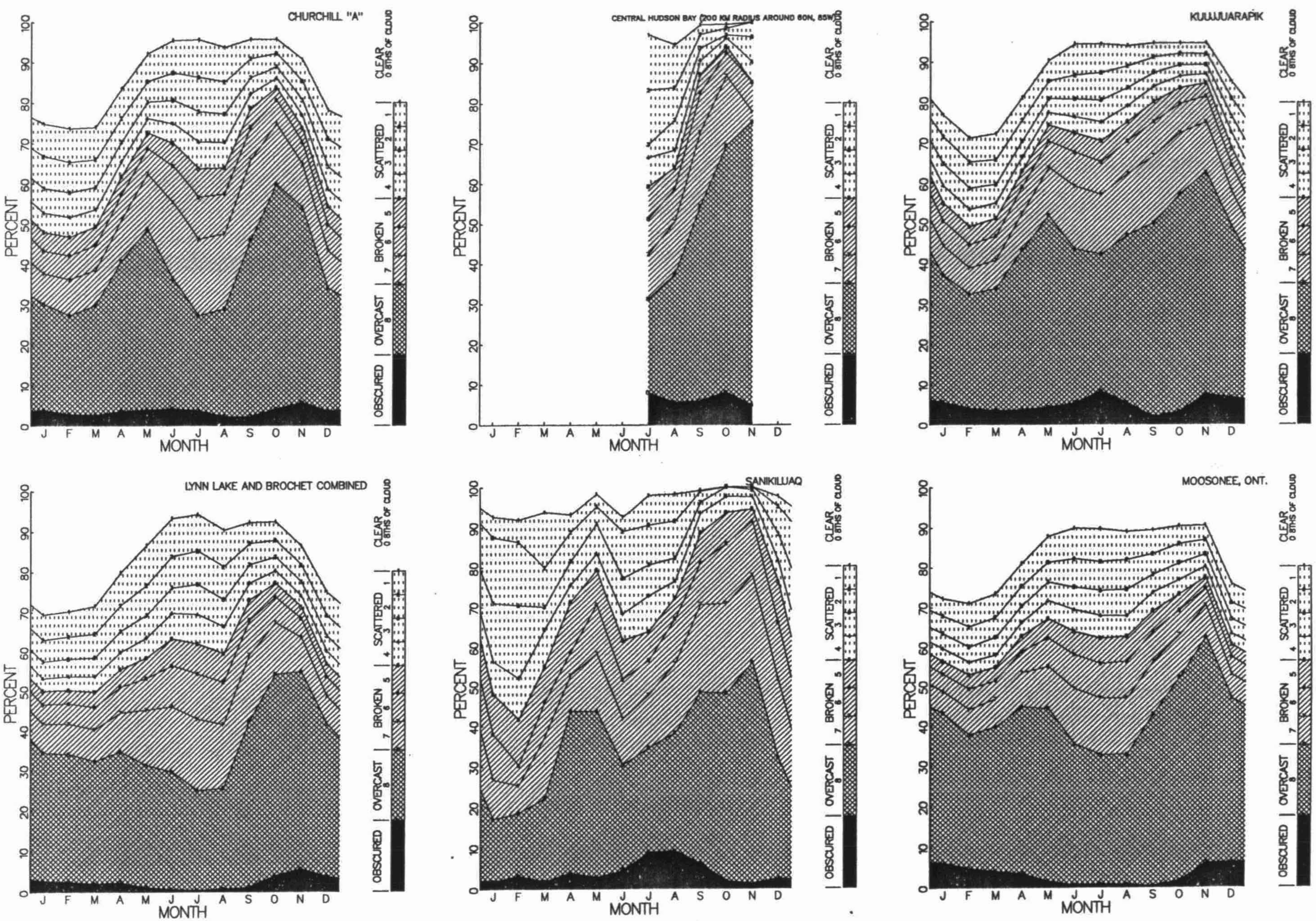
differences are due to milder temperatures and closer proximity to moisture sources for in southern regions of the Bay, more frequent summer storms tracking across the south and eastern side of the Bay, enhancement of precipitation as air travelling across the Bay in summmer picks up moisture and deposits it on the eastern side of Hudson Bay.

The Bay goes through four distinct seasons--winter, spring, summer, and fall--although the timing and length of each season is different than in more southern regions of Canada. Spring occurs during May and June and is characterized by increasing solar radiation, frequent above freezing temperatures, and increasing cloud cover and moisture availability. Snow cover over both land and sea ice melts producing very wet surface terrain. Coastal land fast ice begins to decay and shoreline leads expose increasing amounts of open water. The steady but rapid increase in cloud cover especially in the vicinity of the Bay (*Figure 1.3.4*) is associated with 1) warmer temperatures which increase moisture holding capacity of the air; 2) increased frequency of storms which tend to move across the Bay rather than remaining further to the south as in winter; and 3) increased moisture availability from water logged surfaces. Fog starts to become a serious problem at coastal locations. Precipitation is still quite low ranging from 20 to 40 mm (see *Figure 1.3.4*).

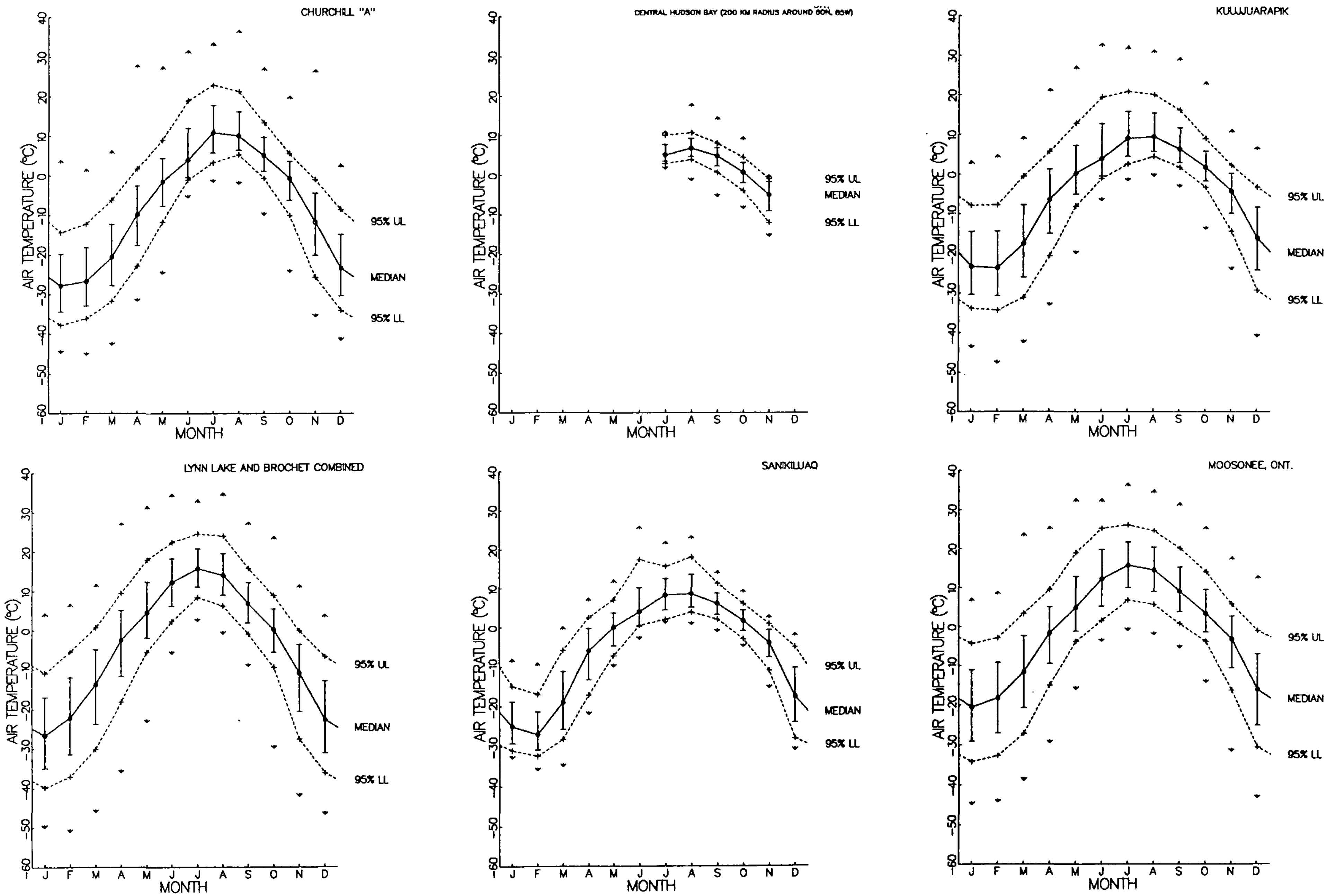
Summer occurs during July and August and is characterized by ice free open water, maximum solar radiation and coastal land temperatures with means of 10 to 15°C and extremes in the mid 20's (*Figure 1.3.5*). This warmer air, however, is considerably modified by the cold waters of the Bay which never reach above 5 to 7°C. July and August are the months with the largest monthly precipitation averaging about 60 mm on the west side of the Bay and just under 100 mm on the east side and in more southern parts of the Bay. Precipitation increases from the northwest to the southeast. This results from more frequent exposure to warmer moisture laden air and increased storms to the south and east of the Bay. Cloud cover remains high during the summer and winds are moderately strong but variable. Winds are lower than during winter ranging from a mean of about 15 to 20 km/h.

Autumn usually lasts from early September to October and is much the stormiest time of the year. At this time cloud cover and precipitation are still at or near their maximum. Frequent cold arctic air masses advance across the Bay picking up moisture and depositing it on the eastern shore. The highest monthly mean and extreme surface wind speeds occur during this time. Temperatures are frequently below zero.

Winter is cold and long lasting from November to April in southern parts of the Bay, extending into May in the more northern parts. Complete freezeup of the Bay occurs by early December. After that, except for persistent open water shore



1.3.4 Percent frequency of monthly cloud cover in five categories (clear, scattered, broken, overcast, and obscured).



1.3.5 Monthly surface air temperature for selected stations around Hudson Bay.

leads, the Bay shows a very similar snow covered surface as the snow covered tundra which surrounds it. During winter, the persistent high pressure over the Keewatin District to the west and low pressure over the Davis Strait/Labrador Sea to the east ensures persistently cold northwesterly winds. Mean temperatures during the coldest month (January) are -25°C with extremes close to -40°C. The best characterization of the region at this time of the year is one of high windchill and blowing snow.

1.4 Ice Cover

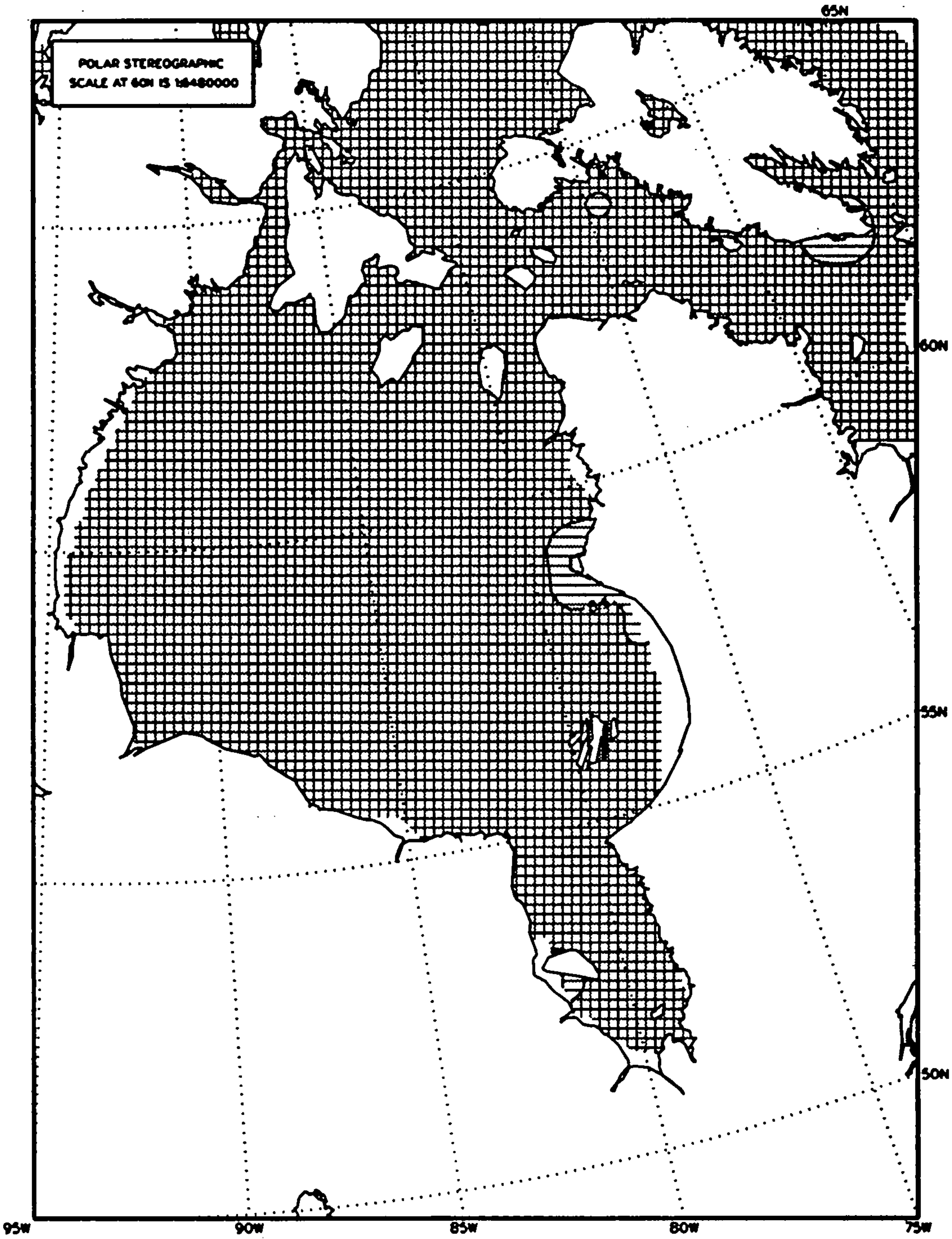
1.4.1 Sea Ice Regime and Sea State

Much of the sea ice material covered here is discussed by Markham (1986 and 1987) and the reader is directed to these papers for more detail. Hudson Bay remains completely ice covered from the beginning of December to the end of May. Despite this long ice season, it does become completely clear of ice almost every year. As a result, almost all the ice is less than one year old.

Before a discussion of the ice regime in the Bay, some basic terminology of sea ice types is required. The initial formation of ice is usually a loose agglomeration of fine crystals called **new ice** which is subdivided based on appearance into grease ice, slush, ice and frazil ice. Once the ice crystals become cemented together into ice with a definite geometry, it is defined as **young ice** and may grow to a thickness of up to 30 cm. Once it is over 30 cm thick, it is termed **first-year ice** which is usually subdivided into thin (30-70), medium (70-120) and thick (over 120 cm). Ice which has survived one or more summers melt is referred to as **second-year ice** or **multi-year ice** if the exact age is not known. The average thickness of multi-year ice is 2 to 5 metres. Multi-year ice is never present in Hudson Bay and second-year ice, advected from Foxe Basin during particularly cold summer and fall seasons, is occasionally present in the northern regions of the Bay. **Pack ice** refers to drifting sea ice caused by surface wind and ocean currents. It is distinct from immobile land fast ice and usually has an ice concentration of seven tenths or more.

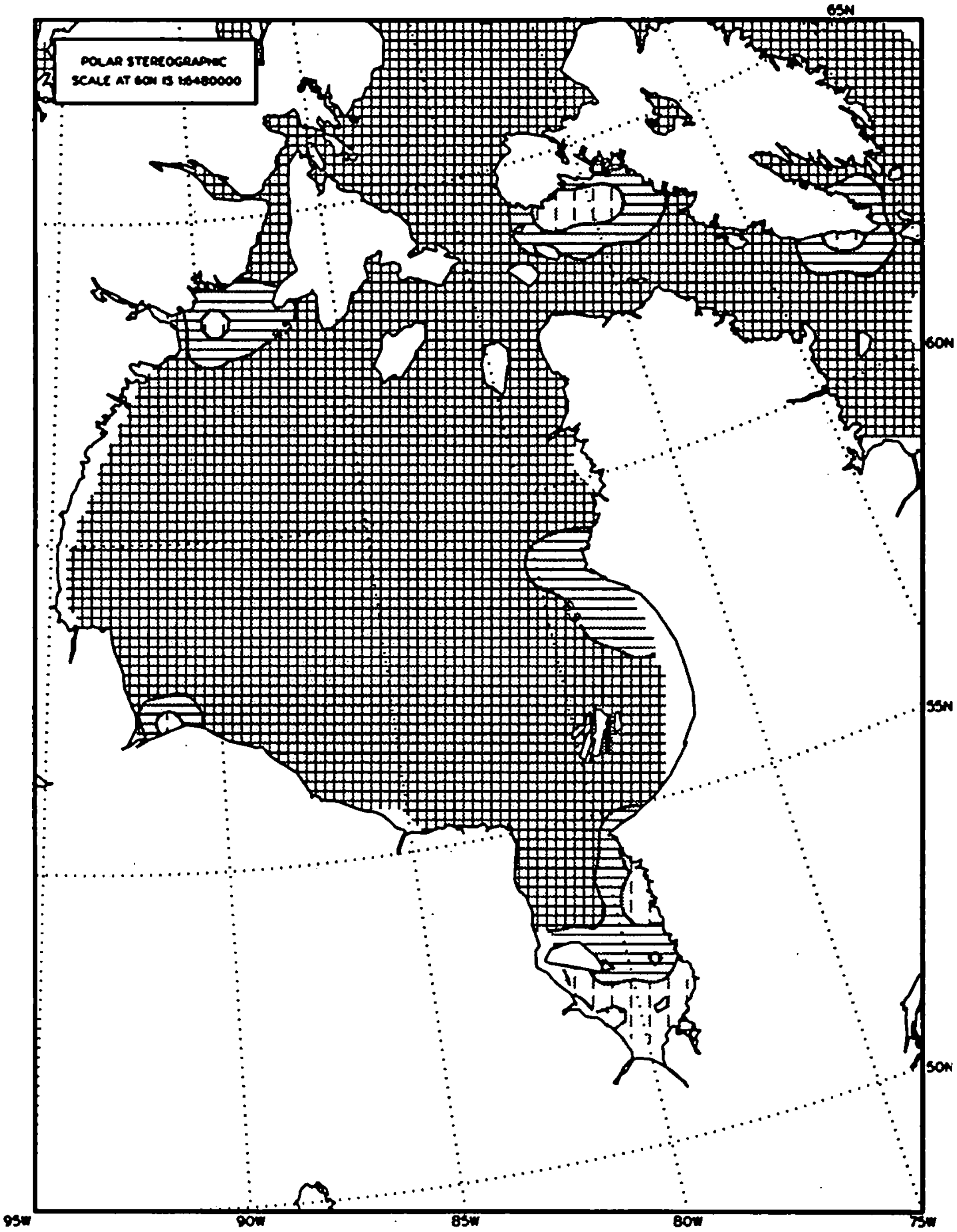
Median patterns of breakup and freezeup are illustrated in *Figures 1.4.1 - 1.4.9*. In early November, freezeup usually begins in the northwestern section of the Bay which is exposed to a high frequency of cold northwesterly winds (*Figures 1.4.7, 1.4.8*). It advances down the west side starting as land fast ice hugging the shore and then advances out into the Bay where it moves southward by counter clockwise surface winds and ocean currents. Freezeup progresses quite rapidly, and by the middle to end of November most of the northern part of the Bay is ice covered and the western part of the Bay is ice covered as far south as the Nelson River. By the first week of December, the

MEDIAN ICE CONCENTRATION
MAY 16 – MAY 31 1972–1988

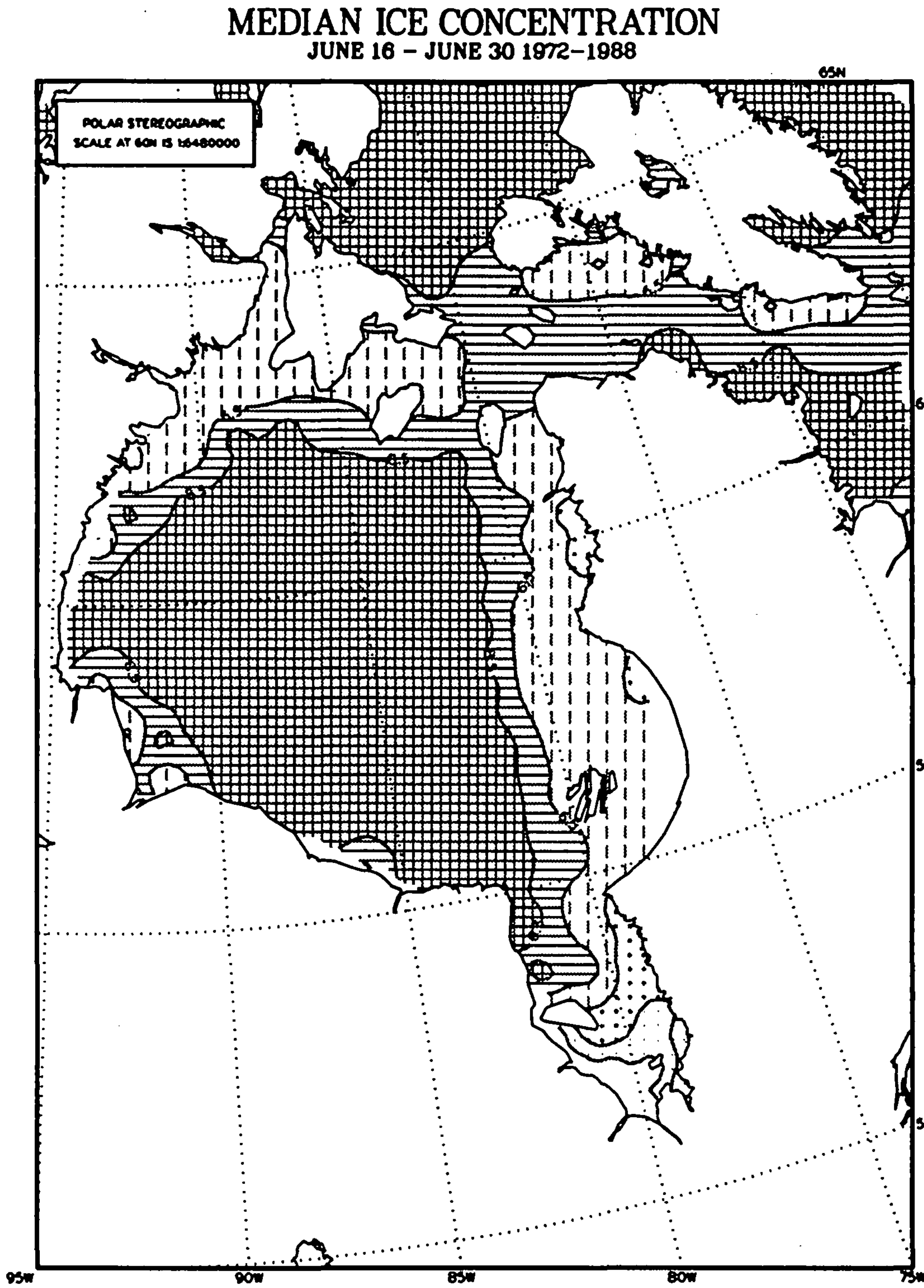


1.4.1 Median sea ice cover for the period May 16 to May 31 based on 1972-1988 data.

MEDIAN ICE CONCENTRATION
JUNE 1 – JUNE 15 1972–1988

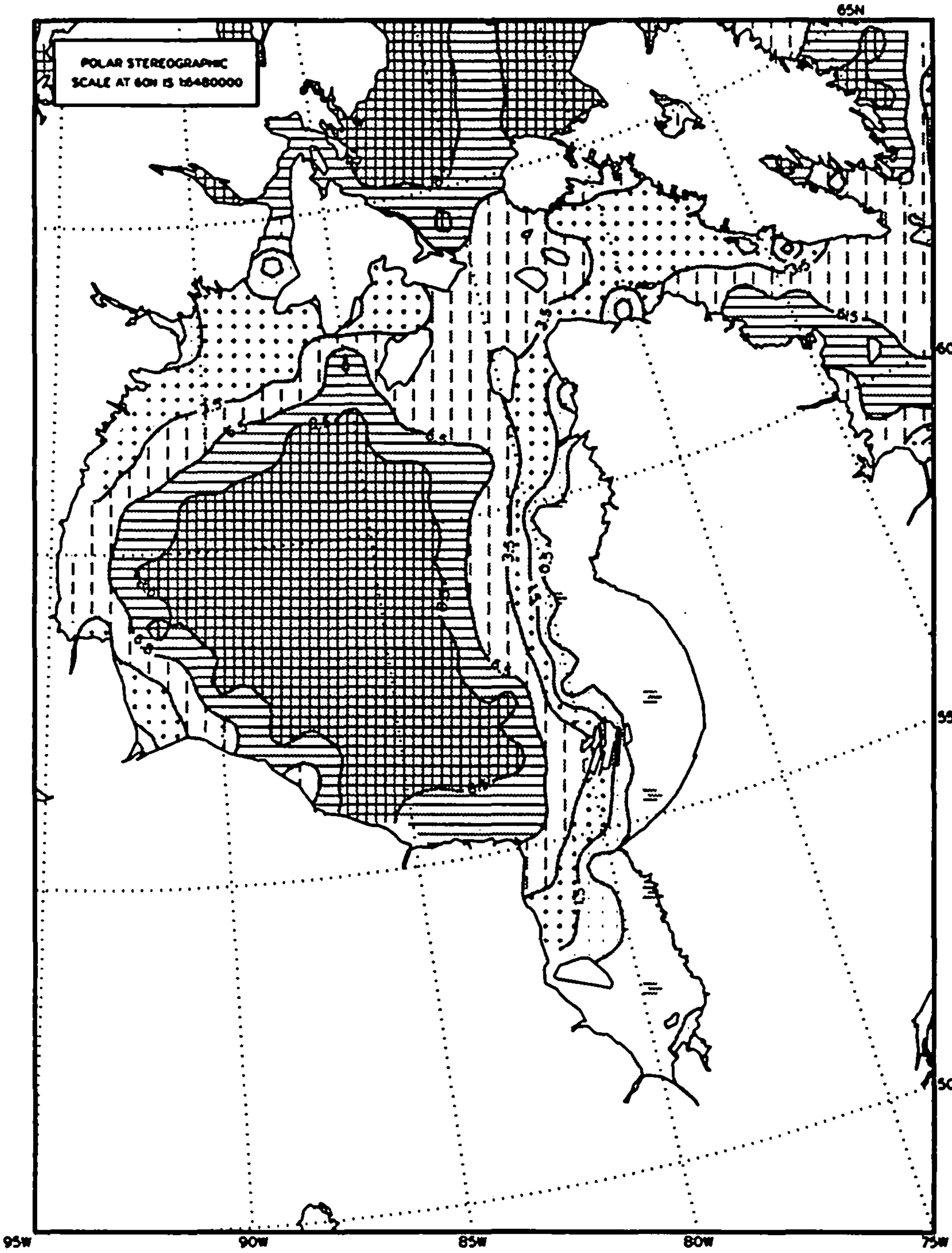


1.4.2 Median sea ice cover for the period June 1-15 based on 1972-1988 data.



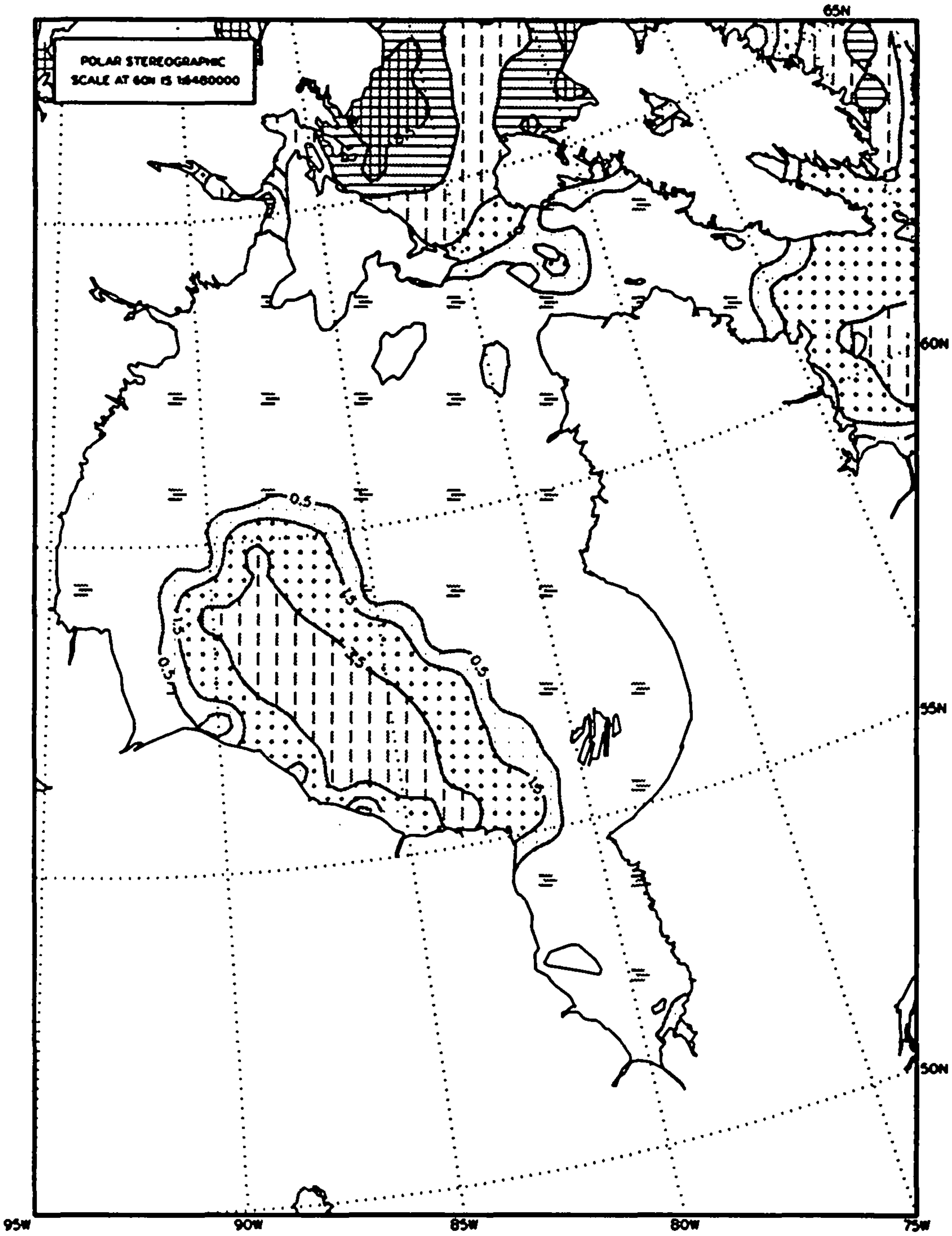
1.4.3 Median sea ice cover for the period June 16-30 based on 1972-1988 data.

MEDIAN ICE CONCENTRATION
JULY 1 – JULY 15 1972–1988

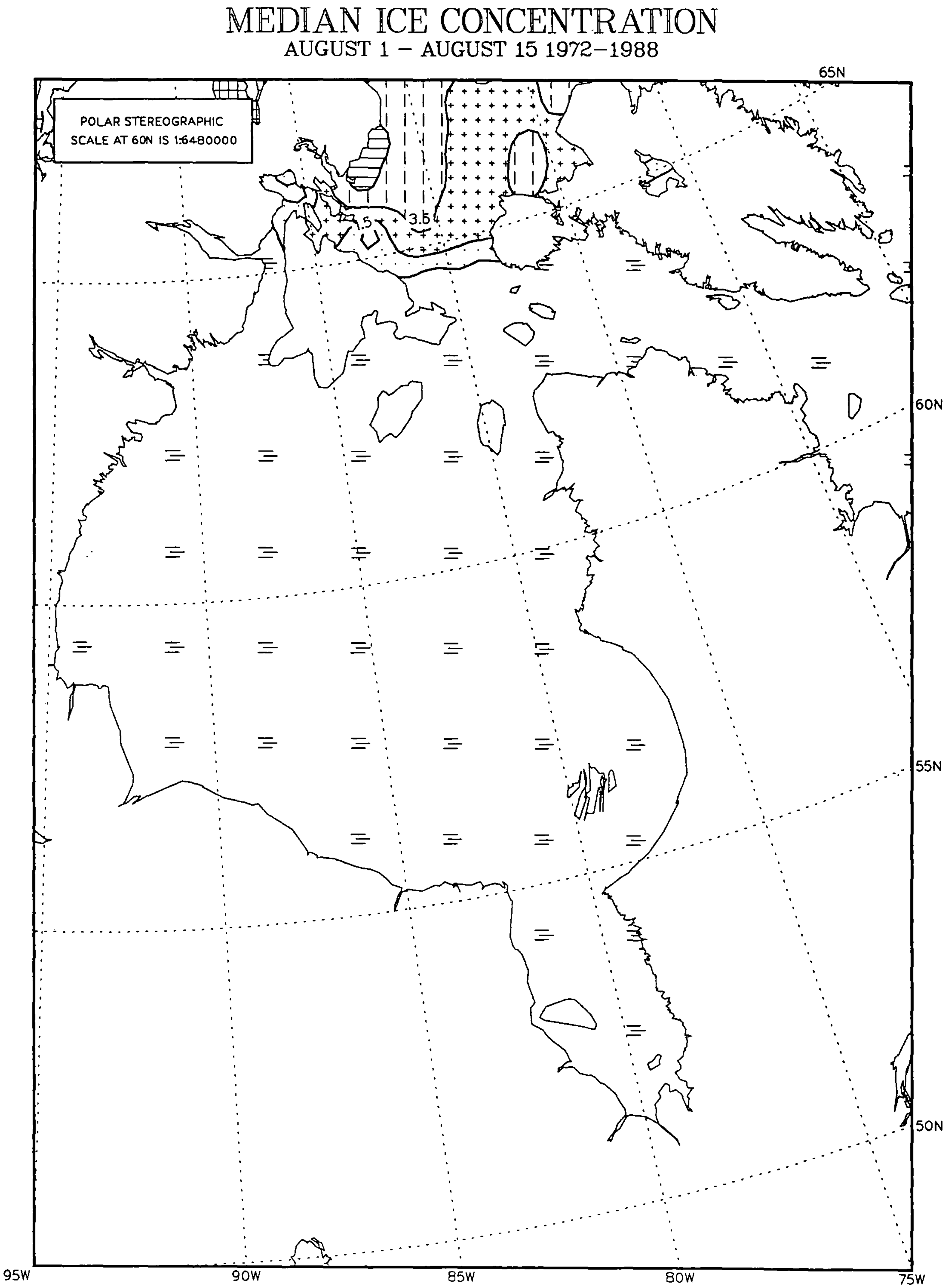


1.4.4 Median sea ice cover for the period July 1-15 based on 1972-1988 data.

MEDIAN ICE CONCENTRATION
JULY 16 – JULY 31 1972–1988

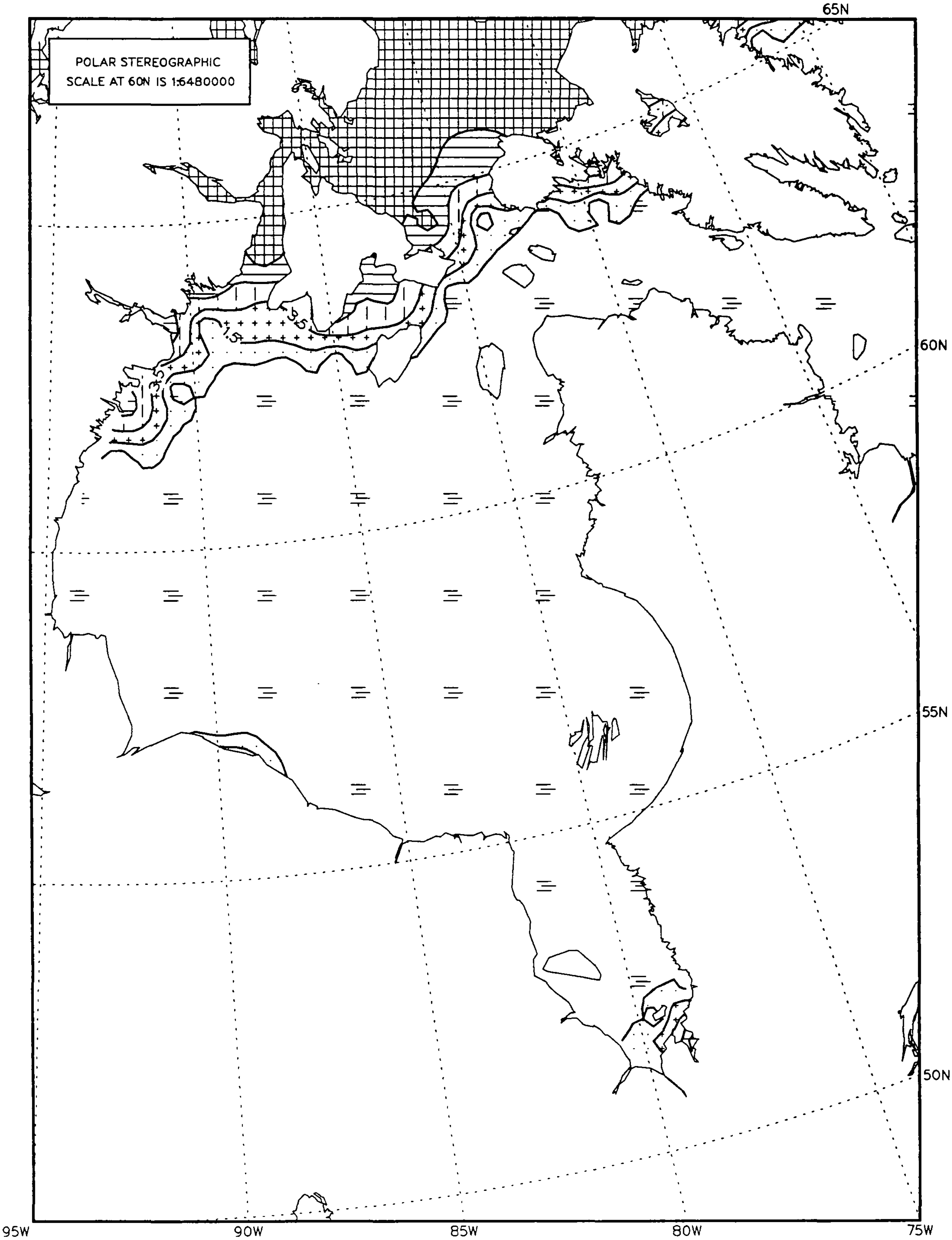


1.4.5 Median sea ice cover for the period July 16-31 based on 1972-1988 data.



1.4.6 Median sea ice cover for the period August 1-15 based on 1972-1988 data.

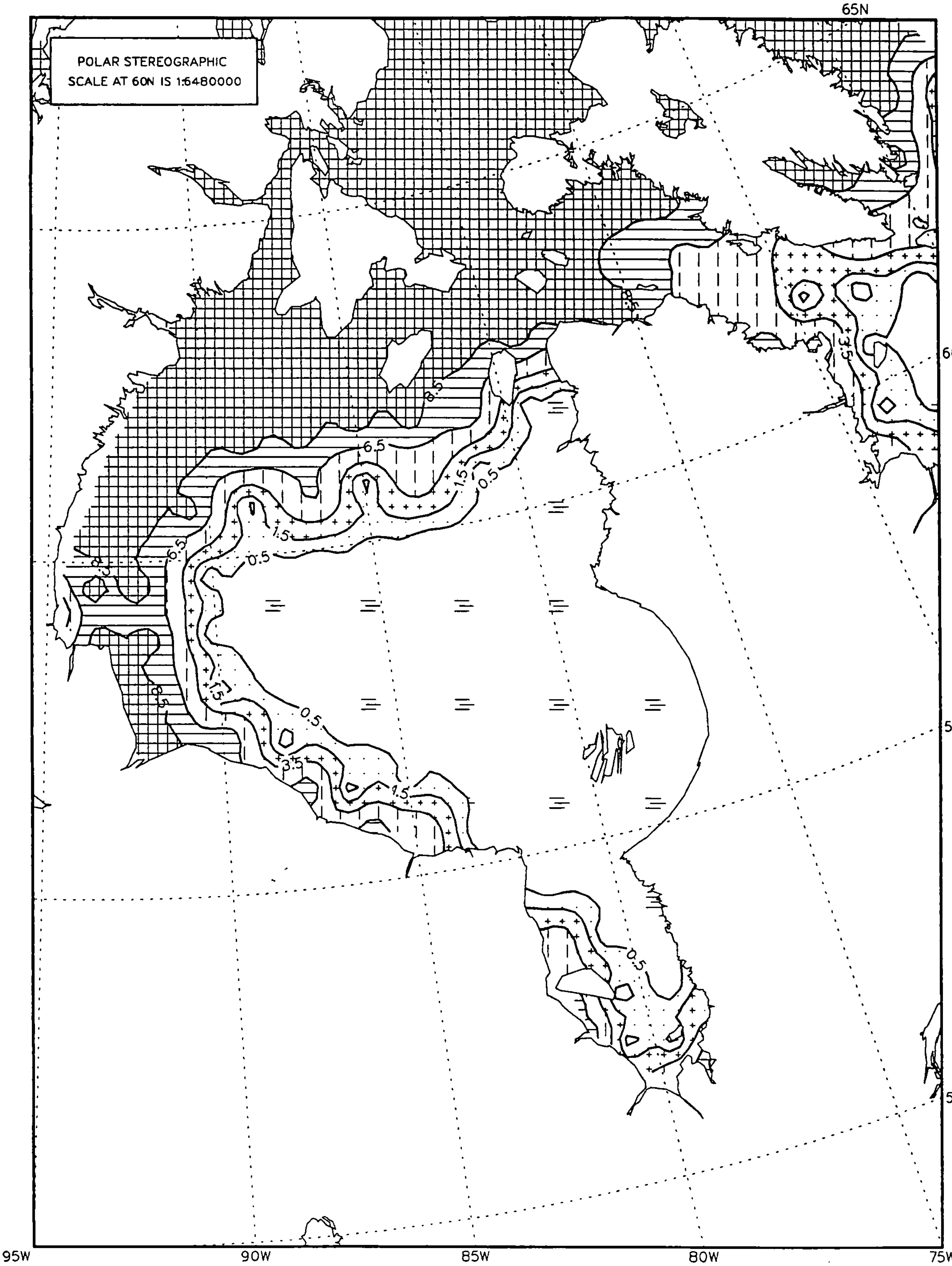
MEDIAN ICE CONCENTRATION
NOVEMBER 1 – NOVEMBER 15 1972–1988



1.4.7 Median sea ice cover for the period November 1-15 based on 1972-1988 data.

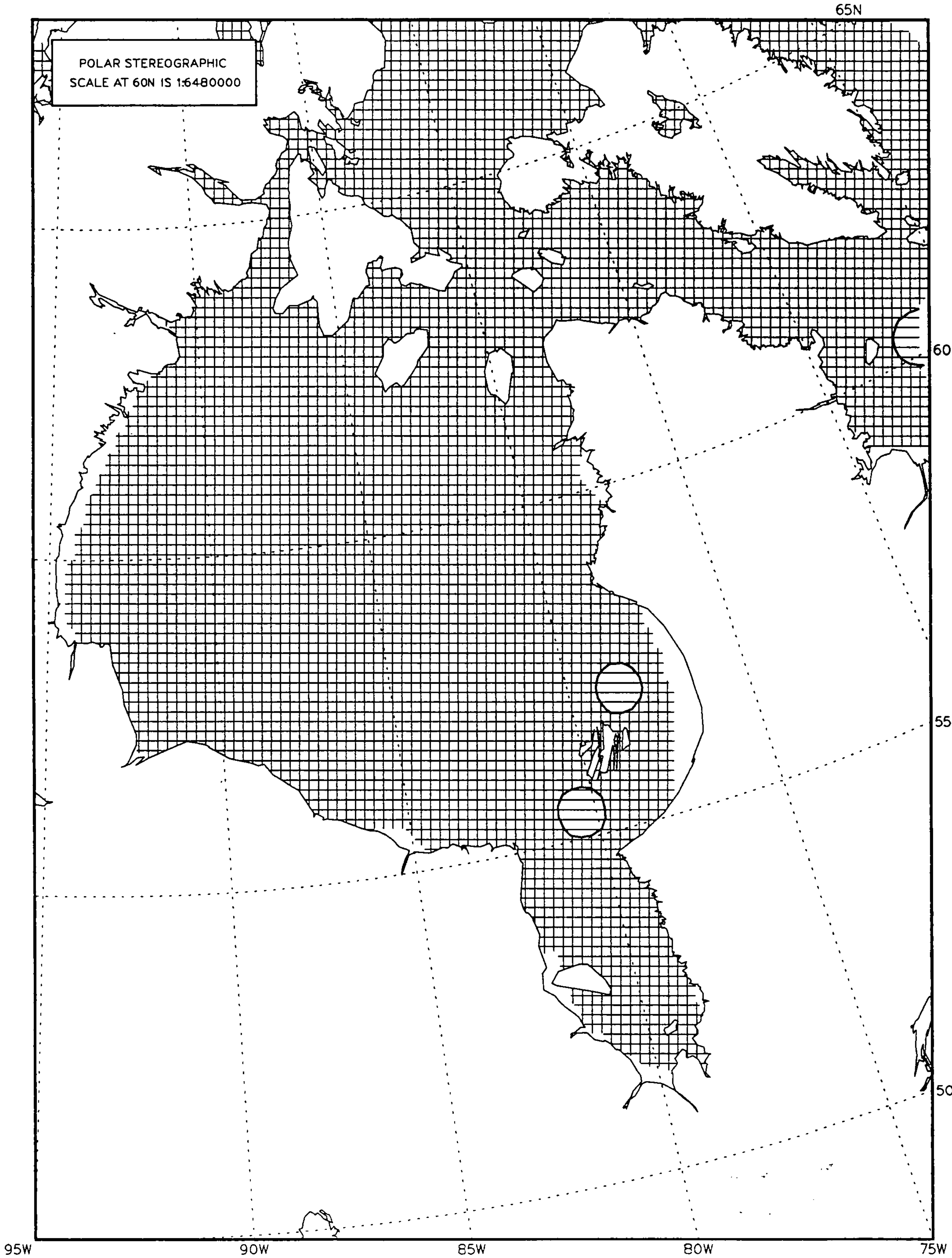
MEDIAN ICE CONCENTRATION

NOVEMBER 16 – NOVEMBER 30 1972–1988



1.4.8 Median sea ice cover for the period November 16-30 based on 1972-1988 data.

MEDIAN ICE CONCENTRATION
DECEMBER 1 – DECEMBER 15 1972–1988



1.4.9 Median sea ice cover for the period December 1-15 based on 1972-1988 data.

Bay has frozen over except for intermittent shore leads. Over the winter, ice cover thickens, progressing through all the stages of young and first-year ice. Away from the land fast ice, the ice is in continual motion. The average displacement is probably several hundred metres per day in a slow counter clockwise direction around the Bay consistent with average wind direction and surface ocean currents.

Continuous ice motion produces ridging and rafting of ice as well as leads which become quickly covered with young ice. Shore leads are most common along the west shore from Churchill to Coral Harbour because of the predominating northwesterly flow. By the end of the ice growth season in April, average thickness of level ice is about 1.6 metres.

In June, with the increased solar radiation, ice begins to decay first by developing a pattern of melt ponds on the sea ice surface from melting snow and ice. By late June, this spreads everywhere over the Bay. Ice clearing usually begins first around Chesterfield Inlet in the northwestern coastal region of Hudson Bay, in James Bay, and along a stretch from Belcher Islands to Mansel Island. In northwestern Hudson Bay clearing is caused mainly by advection as the mean air flow tends to widen existing leads and move the pack ice offshore. Clearing in eastern Hudson Bay and James Bay is caused mainly by northward flow of spring water runoff from rivers which empty into James Bay and Hudson Bay and which accelerates sea ice melt. During July, these three areas expand so that by mid-July pack ice only remains in the south central part of the Bay from Churchill to the Belcher Islands in reduced ice concentration.

The timing of actual ice cover can vary considerably from year to year. *Figure 1.4.10* shows ice concentration at five locations in the Bay throughout the years from 1972 to 1988. At each location, the melt period is longer than the freezeup period and the variability in the length of open water season is evident. In particular, relatively short and long seasons were observed in 1972 and 1981, respectively.

During the open water season significant wave heights over 3 metres occur about 10 percent of the time. This is illustrated with the wave rose (*Figure 1.4.11*), which is for a 200 km radius in the northern part of the Bay. Data were obtained from COADS and Canadian cooperating ships. The predominant direction of the largest waves tends to be from the northwest.

1.4.2 Lake Ice Regime

The Atmospheric Environment Service maintains a database on lake, river, and coastal sea ice conditions within Canada. Observations were initially taken for operational reasons, such

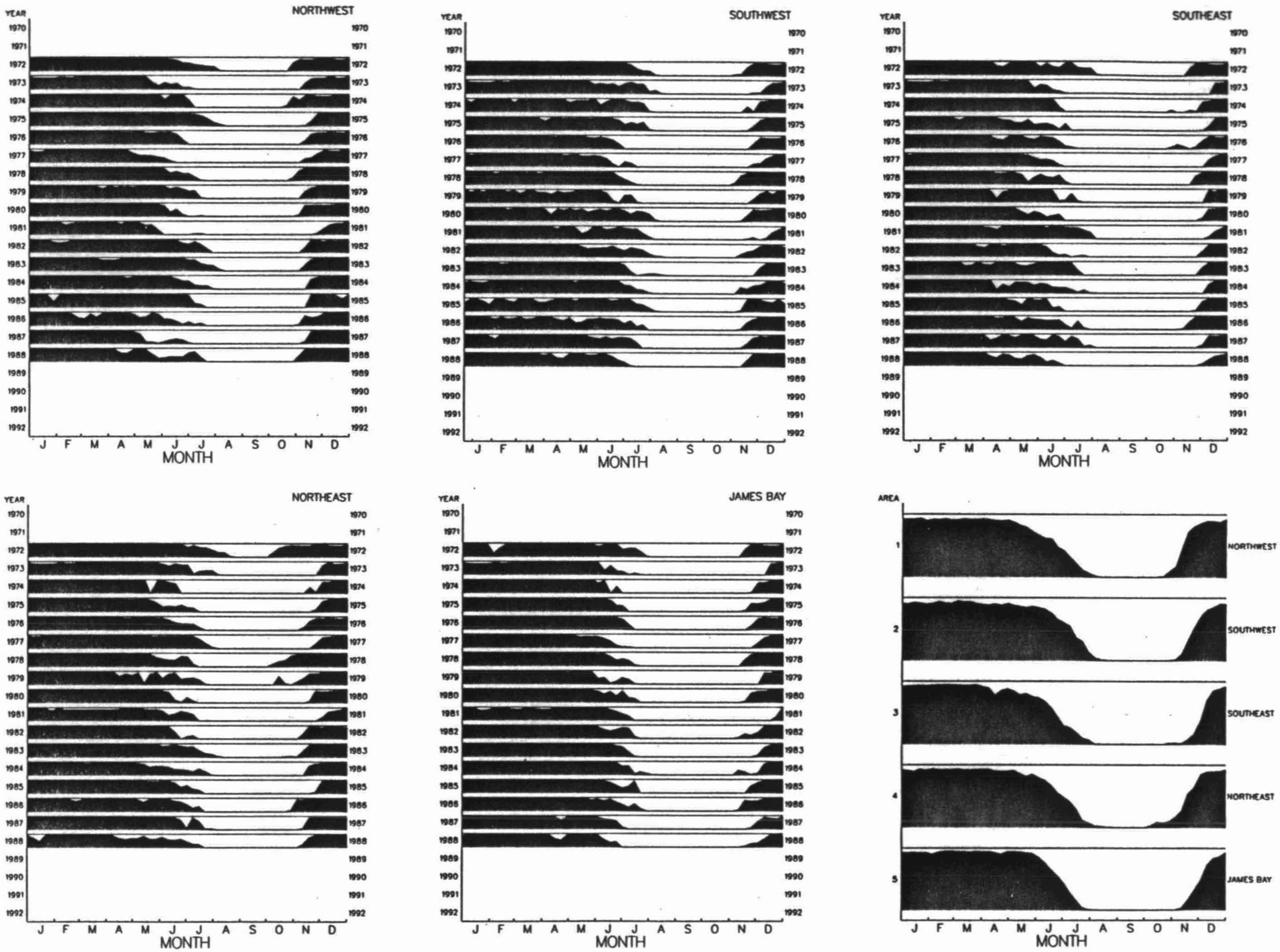
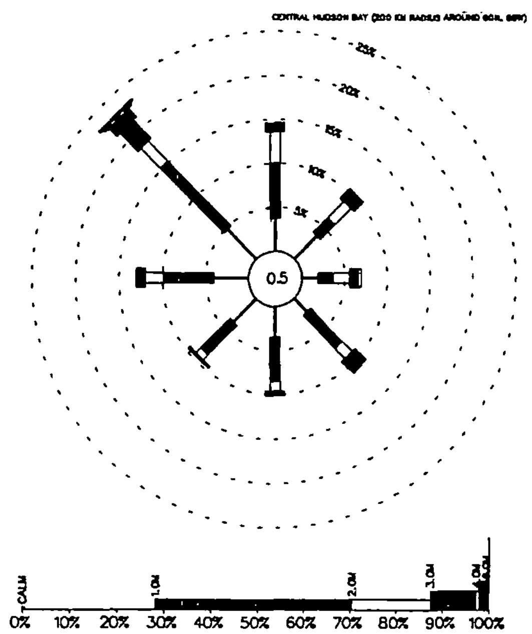


Figure 1.4.10 Mean sea ice concentrations by year and by month for four locations in Hudson Bay (northwest, northeast, southwest southeast) and James Bay.



1.4.11 Wave rose for the summer open water season in the nothern part of Hudson bay based on COADS and Canadian ships-of-opportunity data.

as shipping, but recent efforts have examined ice conditions in the context of climate change (Skinner 1993). The database includes observations for about 250 lakes, either representative of complete small lakes or of a small bay of a larger lake. The information includes the dates of first permanent ice and complete freezeover of ice at freezeup time, and dates of first deterioration of ice and water clear of ice at breakup time. In lake ice studies, the complete freeze over and water clear of ice dates are commonly used because these dates contain less observer bias than the other dates. This reduces the inhomogeneities and thus the variability in the database. The ice event dates are identified and routinely monitored for a number of Canadian lakes. Only a subset of the lakes in the database are used for climate change detection studies.

Air temperature plays a major role in the freezeup and breakup of lake ice. Other meteorological factors that play a role in the formation and decay of lake ice are wind speed and direction, cloud cover, and precipitation. All can affect the final stages of the freezeup and breakup of lake ice cover. Characteristics of lakes such as their fetch, mean and maximum depth, basin geometry, and exposure to wind also determine how the lake will respond to the climate.

The surface water temperature declines in a similar manner as the downward trend in air temperature in the fall. A strong relationship exists between air temperature and freezeup dates. Temperature differences between the two media affect the rate of heat loss from the water to the surrounding air. The loss of heat from the water surface to the air dominates the cooling process. During the freezeup process, ice forms around the shoreline of the lake when the mean daily temperature drops below 0°C. The ice builds outwards as the surface water temperature drops below 0°C. The deepest parts of the lake are the last to freeze. Light to moderate winds speed up the freezing process by removing heat from the water more rapidly and increasing the rate of evaporation. On the other hand, strong winds retard ice formation by breaking up the weak ice that has managed to develop.

The breakup process of a lake is primarily dependent on the temperature regime, the snow cover present on the ice surface, and wind conditions. Snow cover can retard the breakup of ice as it must first melt before the ice itself can melt. However, a greater snow cover can reduce, through the insulation effect, the maximum ice thickness over winter and thus the amount of ice to melt in spring. Once the snow has melted, the reflective properties of the lake surface change allowing an increased amount of solar radiation to be absorbed at the surface. If thawing temperatures are high enough, the ice weakens and begins to melt. The wind acts to free the ice from the shoreline and mechanically break it up.

Lake ice conditions, such as freezeup, breakup, and consequent ice season duration, are dependent on climate variables such as air temperature, solar radiation, wind conditions, precipitation, and snow cover. These conditions can be used as indicators of regional changes in climate. Changes in the mean value of an ice cover parameter may be used to estimate the magnitude of past changes in climate, and to assist in the verification of climate changes indicated by conventional meteorological observations.

Figures 1.4.12 and *1.4.13* show the mean dates for freezeup and breakup, respectively, in the Hudson Bay region. The pattern for freezeup shows the lakes generally freeze earlier in the north and progressively later southward. Typically, freeze-up occurs in early October in the far north and in early November in the southern areas of the region. The breakup pattern is the opposite, and occurs over a longer period than does freezeup. In general, lakes in the south begin to breakup first and the progression is northward. Typically, breakup occurs in mid- to late May in the extreme south and not until late July in much of the north.

1.5 Permafrost

The terrain surrounding Hudson Bay and James Bay is underlain by permafrost. Three distinctive classifications have been identified in the area: continuous, widespread discontinuous, and scattered discontinuous. These distributions are shown in *Figure 1.5.1* (Johnstone 1981). The continuous classification is assigned to areas where permafrost is present everywhere throughout the entire year except for shallow portions near the surface which melt during the summer (the 'active layer'). In this zone, annual mean ground temperature remains below 0°C, and the depth of permafrost can be several hundred metres. In the discontinuous zone, permafrost is not spatially continuous and pockets of unfrozen material are usually present. James Bay is surrounded by discontinuous permafrost while Hudson Bay to the north is surrounded by continuous permafrost. Most of the Hudson Bay wetlands are located within the discontinuous zone (*Figure 1.5.2*; also, see section 1.7).

1.5.1 Churchill Permafrost Station Data

Summaries of ground temperatures taken by the National Research Council of Canada from 1973 to 1983, and by the Canadian Climate Centre of AES from 1985 to present, are used to determine the present state of the permafrost. The thermocouple cable is installed in a rock outcrop covered by thin pockets of moss. The rock was identified as quartzite during the drilling process. The thermocouples are located at 0.5, 1.2, 2.7, 3.5, 4.3, 5.8, 7.3,

Figure 1.4.12 Mean dates of freeze-up for freshwater lakes.

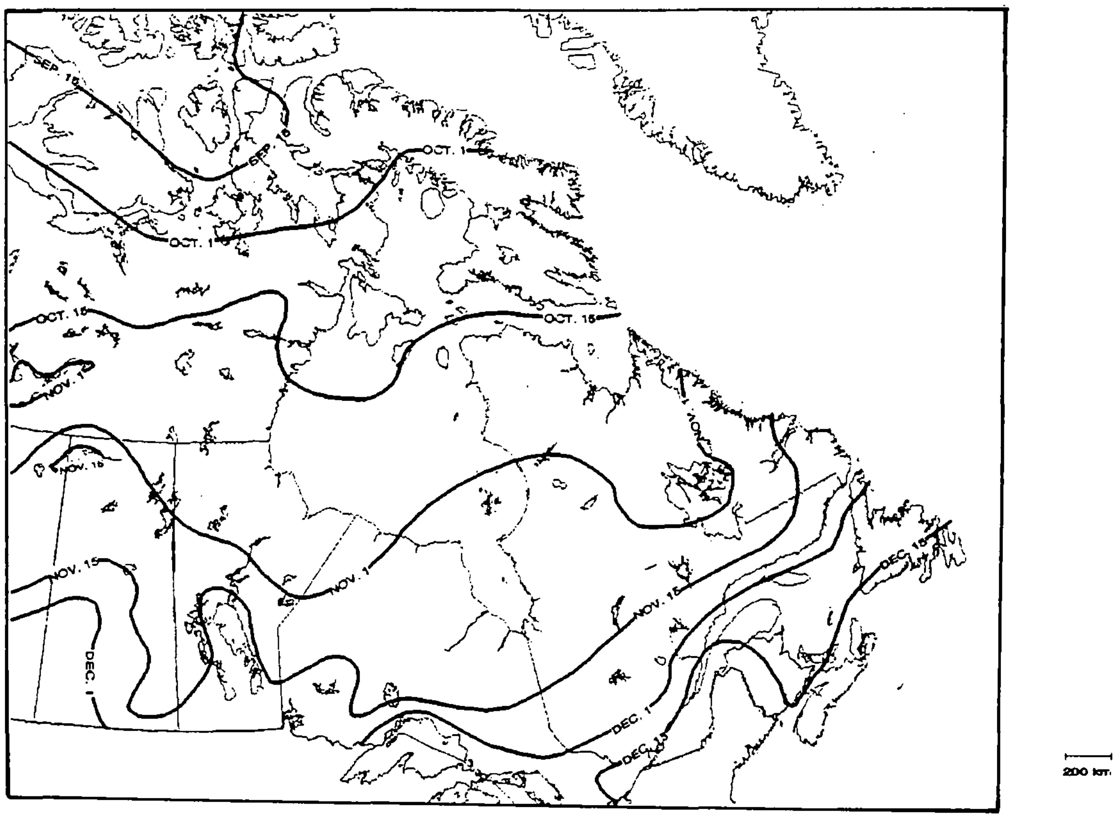
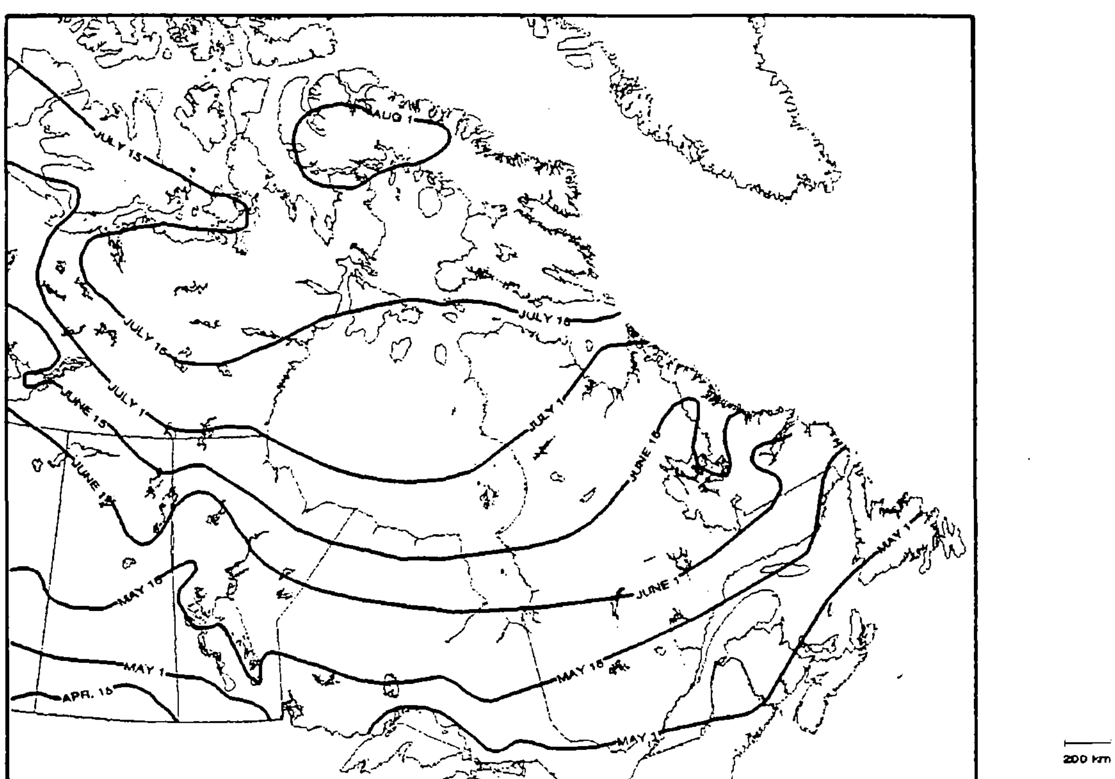


Figure 1.4.13 Mean dates of break-up for freshwater lakes.



Source: Hydrological Atlas of Canada

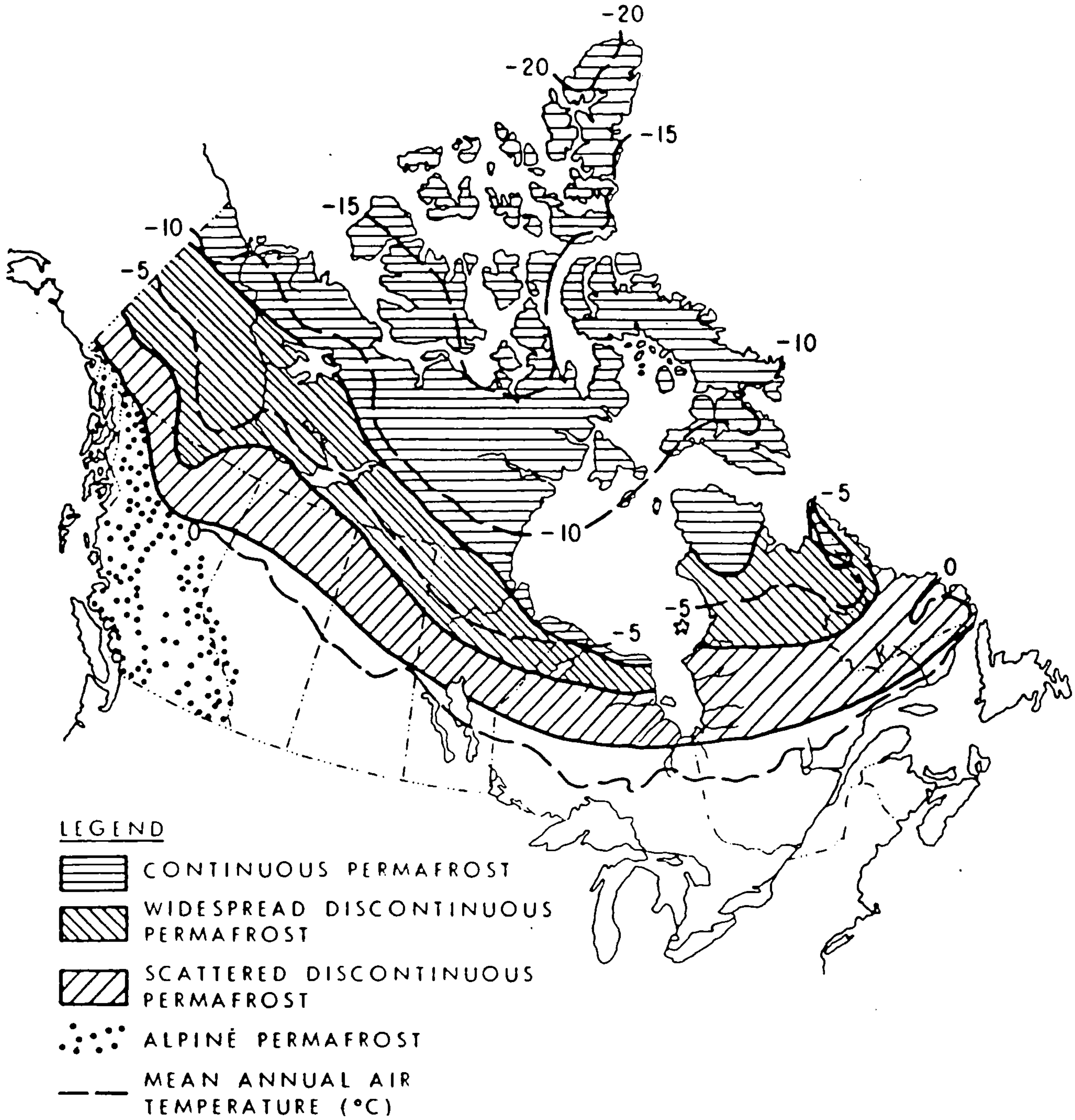


Figure 1.5.1 Distribution of permafrost in Canada (from Johnston, 1981).

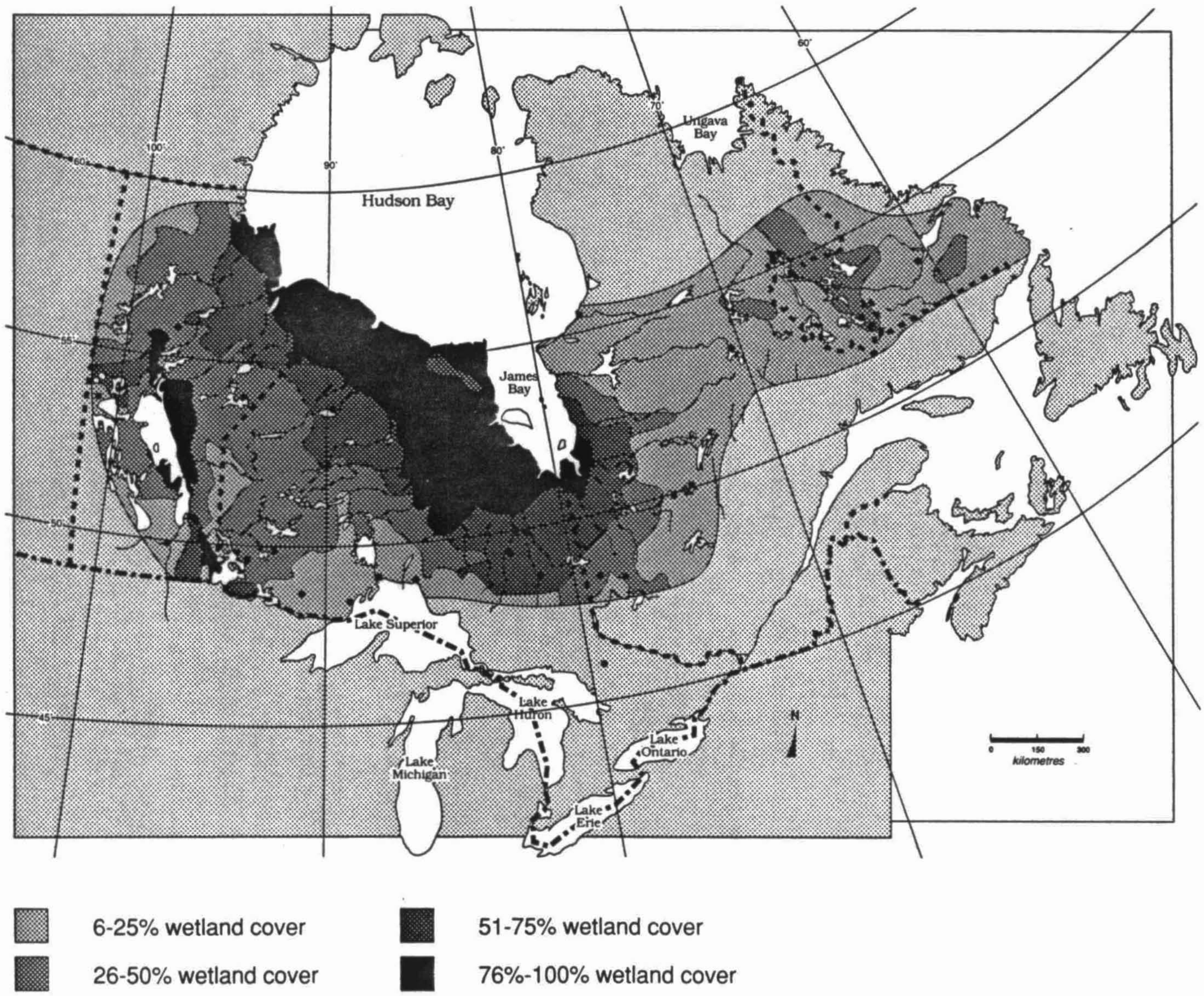


Figure 1.5.2 Wetlands in the Hudson Bay region.

8.8, 10.4, 11.9, and 14.9 metres below the surface. Mean temperature profiles of 12 readings taken in 1978 between Julian days 23 and 348 (1 = January 1, current year), 11 readings in 1983 between Julian days 50 and 350, and readings between Julian day 26, 1988, and Julian day 64, 1989, were computed and are shown in Appendix 1.5.1. All mean temperatures for each recorded level remained below 0°C, indicating the presence of permafrost.

1.5.2 AES Soil Temperature Data

The Churchill permafrost ground temperature data were supplemented with morning soil temperature and snow depth measurements recorded at AES weather stations at Baker Lake, Kuujjuaq, Pickle Lake, and Thompson. The mean temperatures for each observed depth remained below 0°C at both Baker Lake and Kuujjuaq but were positive at Thompson and Pickle Lake. The mean temperature for each depth observed is as follows:

	Baker Lake	Thompson	Pickle Lake	Kuujjuaq
1 cm	-11.0	missing	missing	missing
5 cm	-8.9	3.1	5.4	-1.7
10 cm	-9.3	3.5	5.5	-2.0
20 cm	-8.9	3.5	5.9	-1.2
150 cm	-7.1	3.5	6.2	-0.9
300 cm	missing	2.8	6.0	missing

The positive mean temperatures at both Thompson and Pickle Lake indicate the absence of permafrost at these locations.

The extent to which the permafrost will be affected by climate warming is governed by the vegetation cover, the properties of the soil, and the snow cover (Smith 1989). Snow on the ground is one of the most difficult variables to observe in the north as it collects in low lying areas, near structures, and around vegetation, but is blown clear in open terrain. Snow provides an insulating blanket for the ground, and in areas with a deep snow cover the ground temperatures tend to be warmer. A heavy blanket of snow will reduce the penetration of freezing temperatures into the ground, and in areas with high snowfall permafrost is usually non-existent or its presence is very scattered.

The mean and extreme snow depth measurements for Baker Lake, Thompson, Pickle Lake, and Kuujjuaq are presented in Appendix 1.5.2. They should not be considered as representative of the whole study area; however, the data do show that Thompson and Pickle Lake have more snow cover than the other sites. The mean, standard deviation, and extreme soil temperatures for each station are shown in Appendix 1.5.3. Note that the amplitude of the seasonal cycle is reduced with depth, and the colder stations with less snow cover exhibit larger seasonal changes.

The physical and mechanical properties of permafrost are also temperature dependent. The creep properties of frozen soils and ice are known to be extremely affected by temperature changes, especially when the permafrost temperature is within 1 or 2°C of freezing (Esch and Osterkamp 1989). Although mean temperatures at Kuujjuac are below freezing, they fall within the 1-2°C range and will be affected by any warming of the climate.

1.6 Runoff and Streamflow

The Hudson Bay Bioregion lies within the Hudson Bay drainage basin, one of five ocean drainage basins in Canada. It is the largest both in area with 4.01 million km² (including U.S. portion) and in mean annual flow with a rate of 30,594 m³/sec--draining into Hudson Bay, James Bay, and Ungava Bay. Table 1.6.1 lists the major sub-basins in the Hudson Bay basin with a minimum mean annual discharge of 350 m³/sec at their outlet to the sea.

Table 1.6.1 Hudson Bay Drainage Basin

Sub-basins (with > 350 m³/sec)	Area (km²)	Mean Annual Flow (m³/sec)
Thelon/Kazan	142,400	1,370
Churchill	281,300	375
Nelson/Churchill	1,072,300	3,030
Hayes	108,000	694
Severn	102,800	722
Winisk	67,300	694
Attawapiskat	50,500	626
Albany	135,200	1,420
Moose	108,500	1,440
Harricanaw	29,300	473
Nottaway	65,800	1,130
Rupert	43,300	878
La Grande Rivière/Eastman	97,600	3,400
Grande Rivière de la Baleine	42,700	665
Arnaud	49,500	654
Feuilles	42,500	575
Koksoak	133,400	1,620
Baleine	31,900	581
George	41,700	881
Tha-Anne/Thlewiaza	63,100	507
Seal	50,000	365
Broadback	20,800	382

(Source: Surveys and Systems Information Branch, Environment Canada)

The spatial and temporal variablilty of the freshwater in the basin are represented by two surface water variables, runoff and streamflow. Runoff is the excess from precipitation and snowmelt that is not offset by evaporation, infiltration into the soil or stored on the surface in lakes and bogs or as ice. The amount of runoff is dependent on the amount and intensity of precipitation, rate of snowmelt, permeability of the soil, antecedent soil moisture, surface storage capacity, and evapotranspiration. Runoff collects into channels to become streamflow which increases in volume as it flows through the basin. In addition to surface runoff from the basin land area, streamflow also consists of water derived from direct precipitation into the lakes and channels as well as a component of groundwater flow.

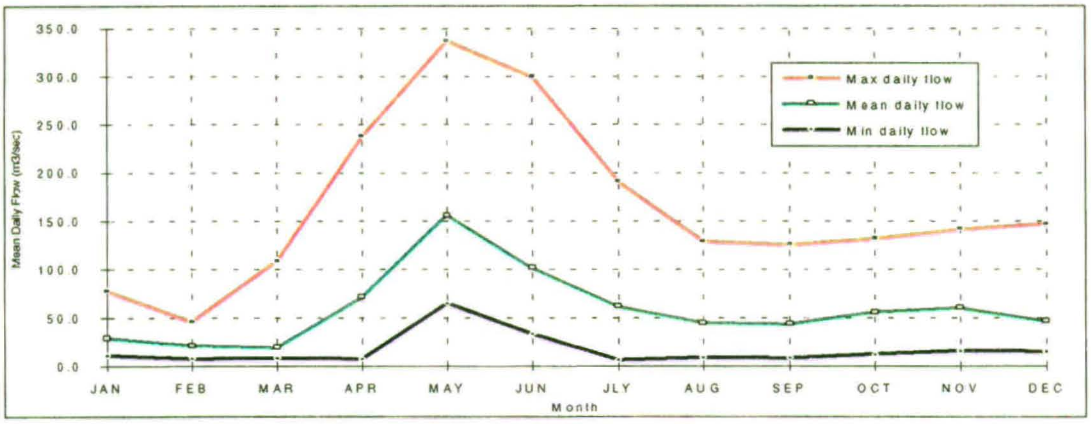
The basic geological formations of a region with their superimposed topographic features play a vital role in the initiation and subsequent distribution of precipitation, the gathering of the resultant runoff into river systems, and the recharging of groundwater flow systems. The Hudson Bay drainage basin lies primarily in the Canadian Shield geological region. Parts of the Nelson and Churchill sub-basins lie in the Great Plains region but that portion of the basin is not of interest to this study. The topography of the Canadian Shield is confused with numerous lakes and wetlands scattered throughout the region and the terrain is generally impervious, with ample but irregular gradients. Its climate, as noted earlier, varies from sub-arctic to arctic with average annual precipitation increasing from the northwest to the southeast and varying from a low of 250 mm in the extreme northwest to a high of about 1,000 mm in southern Labrador. In summer, evaporation from lake areas can be substantial but the sparseness of vegetation and imperviousness of the rock structure contribute to a high rate of surface runoff which increases from west to east ranging from 150 mm to 700 mm. *Figure 1.6.1* shows the mean monthly flow of three streamflow stations in the bioregion. They illustrate the seasonal variability of streamflow within the basin which is characterized by a single peak flow resulting from the simultaneous release of water stored as snow and that held frozen in rivers. The timing of the peak flow generally occurs from May to August varying with the station latitude in response to the northward progression of snow melt and the breakup of lake and river ice.

1.7 Wetlands

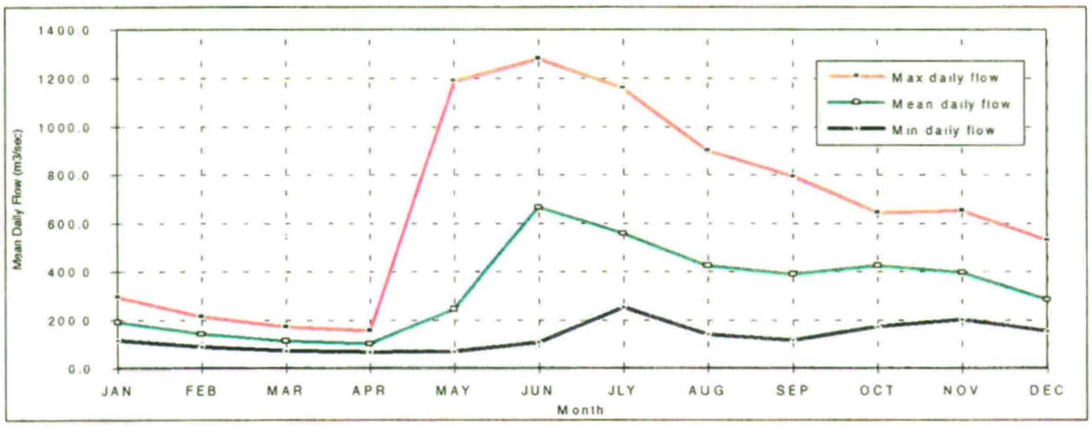
Wetlands are lands saturated by surface or near surface waters for periods long enough to promote the development of hydrophytic vegetation and gleyed or peaty soils. They provide a habitat for a wide range of plants and animals, and play a critical role in the hydrological regime of a region affecting water purification, groundwater discharge, and flood peak modification and in making

Figure 1.6.1 Mean Monthly Flow

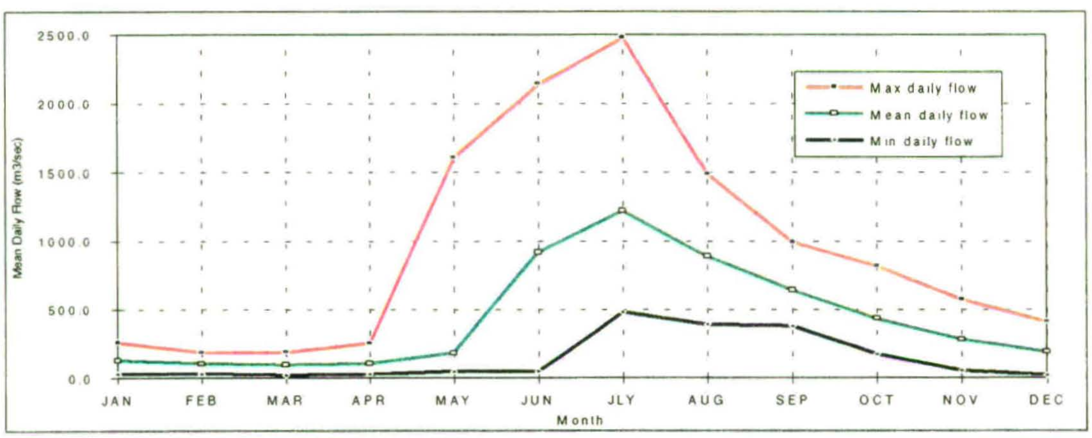
(a) 04NA001 -- Harricana (Riviere) A Amos
Lat: 48°36'02" N Long: 78°06'34" W Area: 3680 Km^2



(b) 03EA001 -- Baleine a la sortie du lac Bienville
Lat: 54°50'57" N Long: 73°59'01" W Area: 21,000 Km^2



(c) 06LC001 -- Kazan River above Kazan Falls
Lat: 63°39'10" N Long: 95°51'19" W Area: 70,000 Km^2



development environmentally sustainable. Because wetlands are known to produce both carbon dioxide and methane, their role in the global cycling of these greenhouse gases is of particular interest to current concerns on global warming.

A comprehensive research program, the Northern Wetlands Study (NOWES), was developed to determine the role of Canadian wetlands in affecting the physical and chemical change in the environment (Schiff and Barrie 1988). NOWES focussed on those areas in the provinces of Manitoba, Ontario, Quebec, and Newfoundland (Labrador) where wetlands form a significant component of the land surface. *Figure 1.5.2* shows the distribution of wetlands within the NOWES study area which forms a large portion of the Hudson Bay Bioregion study area. Mortsch (1990) has compiled a comprehensive resource document providing details on the socio-economic, air chemistry and climate, physiography and geology, soils and vegetation, and hydrology for the eastern Canadian boreal and sub-arctic wetlands. The following discussion on the hydrological characteristics of northern wetlands has been extracted from that resource document.

The role of wetlands within the hydrological cycle was once considered to be storage and low-flow (base flow) augmentation for surface streams and/or groundwater. However, current research suggests that many more wetlands are groundwater discharge areas than recharge sources (Carter and Novitzki 1988). The hydrological role of wetlands is very complex and site-specific, thus, generalizations based on physiographic setting cannot be made. Carter and Novitzki (1988) identified specific factors which affected wetland hydrology:

1) the inter relationships between local, intermediate, and regional groundwater tables;

2) position of the local water table;

3) geologic setting;

4) ratio of vertical to horizontal hydraulic conductivity (a function of wetland size, slope, peat accumulation, and presence or absence of permafrost);

5) depth and width of the basin;

6) local slope and relief; and

7) the location of groundwater divides.

The streamflow regime of basins in the northern latitudes of eastern North America containing wetlands has been investigated by Price and Woo (1988) and Woo and Heron (1987a) in the Sub-Arctic and Roulet and Woo (1986, 1988) in the Arctic. The major control on the basin hydrograph is the relationship between the wetland and the regional groundwater system and the saturation state of the wetland prior to events such as snowmelt and summer storms. In more northerly latitudes these factors, along with the duration of seasonal frost (especially in sub-arctic wetlands) and the presence of permafrost, are important.

Conditions which limit the storage capacity of the wetland tend to favour event responsive hydrographs from basins containing wetlands. South of the discontinuous permafrost zone, this is caused primarily by high antecedent saturation of the wetland as a result of a high regional water table and/or previous precipitation events (Taylor and Pierson 1985). Thick peat accumulation provides storage capacity which can attenuate flows of individual storm events but tends to have little influence on surface runoff yields on a seasonal or annual basis (Bay 1967, 1969).

In sub-arctic and arctic regions the presence of late seasonal frost, such as occurs under the more densely forested areas (Cowell et al. 1978, Woo and Heron 1987a), and permafrost also limit the storage capacity of wetlands and enhance event responsive streamflow. Roulet and Woo (1988) noted that, although the presence of organic soil layers in northern wetlands can provide a significant amount of storage capacity, the effect is only temporary. They observed that 70 percent of the total measured basin flow was released within two weeks of the initiation of streamflow in an arctic basin located near Baker Lake, Northwest Territories. Roulet and Woo (1988) concluded that the presence of frozen soils throughout the entire basin reduces the spatial contrast of variable hydrological systems (i.e., mixed wetland and upland systems).

Woo (1988) proposed a 'wetland streamflow regime' for arctic and sub-arctic regions on the basis of these and other studies. This northern wetland flow regime attributes the attenuation of streamflow during the dry season to the presence of a frost table close to the surface and low hydraulic gradients resulting in retention of water in ponds and organic soils. A large spring freshet is produced by abundant snowmelt and enhanced by low infiltration into the frozen ground. The exception is where winter snowfall is so low that the spring snowmelt is insufficient to generate overland flow. In another detailed study of a sub-arctic coastal marsh located in the Hudson Bay Lowland (southern James Bay), Price and Woo (1988) found that snowmelt and rainfall are transmitted quickly through the system in spite of the large total water storage capacity. The only exception occurred following long dry periods when evaporation created a large water deficit.

It is not certain if the same streamflow regime holds for wetland basins with deep peats and minimal upland areas. In the case of large sub-arctic peatland complexes, such as in portions of the continuous wetlands of the Hudson Bay Lowland, it can be postulated that Church's (1974) muskeg regime, in which flood flows within peatlands are greatly attenuated because of the large water-retaining capacity of peatland vegetation and due to the high resistance to runoff presented by the vegetation and its irregular surface, may be more representative. In these areas, particularly outside the zones of continuous or widespread permafrost, the lack of normal surface streams, a regionally low gradient, and thick peat material could provide considerable water retaining capacity even though regional water tables are high. Large inputs to the peatland in the form of rain, snow melt, or contributions from deeper aquifers might result in a change in volume of the peat mass, flooding of the peat surface, and/or even a quasi-floating of the peat mass. In any case, there would likely be a less immediate response in the hydrologic regime of rivers or small streams draining coastal marshes.

Woo and Heron (1987b) investigated the breakup pattern of small sub-arctic rivers along the coast of James and Hudson Bays. They note that the control of breakup varies between temperate and sub-arctic rivers. The breakup of rivers in temperate regions is controlled primarily by river ice whereas in sub-arctic regions, channel snow cover is more important. The presence of large areas of wetlands common to the Sub-arctic is considered to be an important factor affecting breakup due to seasonally frozen ground, flat terrain, and snow and ice conditions which combine to produce complex hydrological responses during the breakup season, particularly in small catchments.

The specific breakup sequence for a small river draining wetlands in the Hudson Bay Lowland was described by Woo and Heron (1987b). They note that snow conditions dominate the breakup sequence until the underlying ice layer is exposed. The early portion of the sequence is characterized by a large amount of snow melt draining from the wetland. Ponds are impounded behind snow dams created by compacted snow in the channel. This accentuates flooding as the dam is breached causing slushing and removal of the residual snow cover. At this point channel ice conditions become most important to the river breakup process.

2. REGIONAL CLIMATE VARIABILITY AND TRENDS

2.1 Nature of Climate Variability

Until the recent concern over climate change, the traditional approach to resource management and planning assumed the climate of a large region such as Hudson Bay to be stable though still naturally varying. The variability of Hudson Bay climate, whether natural or from other causes, has been considerable over the instrumental record. For example, 1972 was a particularly cold year and 1981 was particularly mild. Climatologists have been aware since the early part of this century that variatons of climate at different locations around the globe are not independent but that changes in one region are related over great distances to changes in other regions. These relationships are referred to as teleconnections (Glantz et al. 1991) and result from the wave-like nature of the global climate system. For example, correlations have been found between anomalously cold temperatures over Hudson Bay and drier-than-normal conditions over South America (Hartmann 1984).

By calculating the correlation of climate parameters around the globe, patterns in the variability of climate can be obtained. These patterns are usually associated with major oceans. The best known is ENSO (El Nino-Southern Oscillation) in the tropical Pacific Ocean. For the Hudson Bay area the teleconnection patterns which have the most influence are the North Atlantic Oscillation (NAO) and the Pacific/North American Oscillation (PNA). The NAO is an oscillation in the intensity and position of the Icelandic Low and the Atlantic sub-tropical High. During intensification of the Icelandic Low, the pressure gradient over the Bay strengthens, producing more frequent northerly flow. At the upper level, the 50 kPa trough deepens. The PNA is associated with intensification of the Aleutian Low in the north Pacific, strong ridging over western North America, and a deeper-than-normal trough over eastern Canada.

Feedbacks between different components of the climate system also contribute to climate extremes. The best known feedback is the sea ice/albedo effect (LeDrew et al. 1992) where an increase in temperature will melt more ice and snow. This will reduce surface albedo and increase solar radiation absorption, which will further increase surface temperatures. Other feedbacks involve temperature-moisture-cloud-reflected solar radiation and sea ice thickness-solar energy-sea surface temperature. Work by Prinsenberg (1984) suggests that since melting of the ice cover in spring and early summer takes 50% of the solar energy available and most of the rest is used to increase summer water temperatures, the severity of the preceding winter is inversely related to surface water temperatures the following summer.

2.2 Variability of Maximum and Minimum Temperatures and Total Precipitation

The climate of the Hudson Bay Bioregion is typified by the extreme diversity of its temperature and precipitation distributions - south to north, east to west, and in its seasons. Air temperature has been traditionally analysed in its daily average form, that is, the arithmetic mean of the daily maximum and minimum temperatures. However, this daily swing of temperature extremes in this geographically large region has significant physical importance.

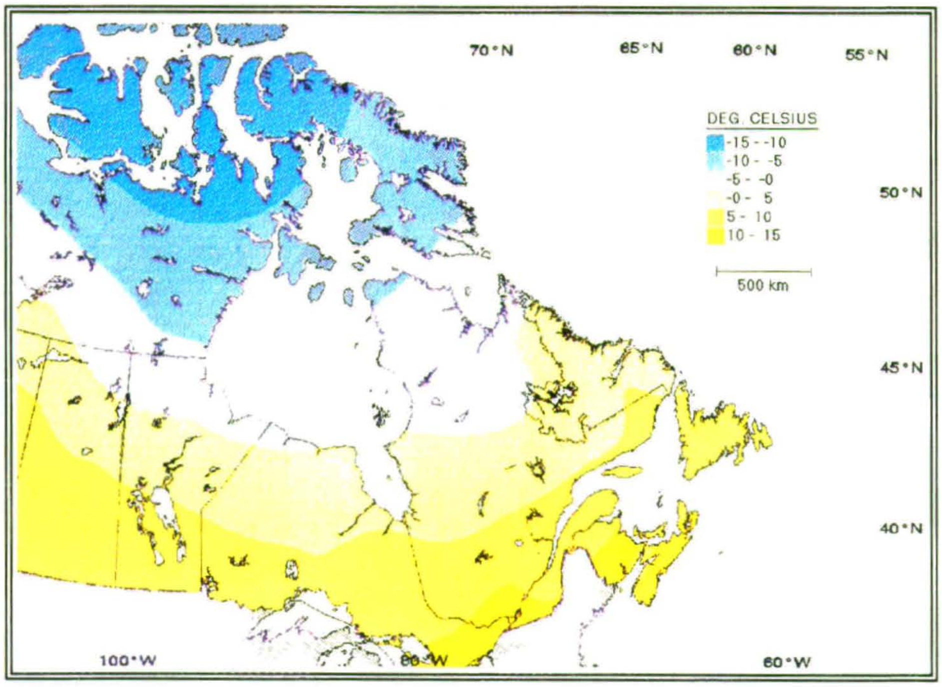
The maximum temperature for the day usually occurs during the afternoon hours, and is therefore thought to be more or less representative of the peak day-time temperatures. On the other hand, the minimum temperature usually occurs in the early morning, pre-dawn hours, just before sunrise, and is therefore usually considered to be most representative of the lowest night-time temperatures. Recent studies have shown that for a large portion of the Northern Hemisphere landmass, including Canada, night-time minimum temperatures are increasing more than are the day-time maximum values during the past 100 years (Karl et al. 1991; Skinner and Gullett 1993).

Figure 2.2.1 shows annual average maximum temperatures vary from about 0°C to 5°C across the southern areas of the region to -5°C to -10°C across the north. The location of the 0°C isotherm of average annual maximum temperatures represents the approximate location of the line of discontinuous permafrost. Also, the location of the -15°C isotherm of average annual maximum temperatures represents the approximate location of the present day tree line.

Extreme variation exists in the range of average temperatures both in time, daily, seasonally, and annually, and in space throughout the bioregion. Figure 2.2.2 shows annual average minimum temperatures vary from -5°C to -10°C across the southern areas of the region and from -15°C to -20°C across the north. Maps showing the distributions of average winter (average of December, January, and February) and summer (average of June, July, and August) season maximum and minimum temperatures are shown in Appendix 2.2.1 to 2.2.4, respectively. Average winter maximum temperatures range from -10°C to -15°C in the south to less than -25°C in the north, while average winter minimum temperatures range from less than -15°C in the south to less than -35°C in the north. Average summer maximum temperatures range from greater than 20°C in the south to about 10°C in the north, while average summer minimum temperatures range from less than 10°C in the south to just greater than 0°C in the north.

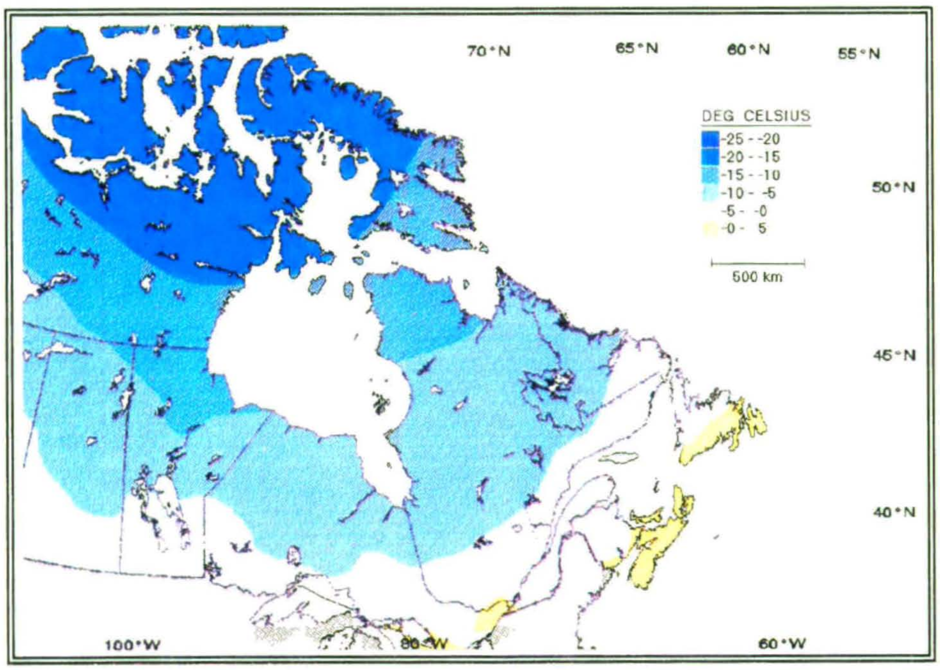
Extreme variation also exists in the range of average total precipitation both in time, seasonally and annually, and in space throughout the bioregion. Figure 2.2.3 shows the distribution of annual average total precipitation. A strong average precipitation gradient exists across the region from northwest to southeast, with extremely low annual totals of generally less than 200 mm a year on the northwestern shores of Hudson Bay in the Keewatin Region of the NWT, to less than 500 mm in the dry boreal forest area of northern Manitoba, to greater than 800 mm

Figure 2.2.1 Annual average maximum temperature, 1951-80.



Source: Environment Canada

Figure 2.2.2 Annual average minimum temperature, 1951-80.



Source: Environment Canada

in the more humid boreal forest area of central Quebec. Maps showing the distributions of average February (representative of the winter season) and August (representative of the summer season) are shown in Appendix 2.2.5 and 2.2.6, respectively. Average February precipitation totals are extremely low, ranging from less than 25 mm in the north and northwest to less than 50 mm in the southeast. Average August totals are generally higher, ranging from lows of 25 mm to 50 mm in the north to greater than 100 mm in the more humid southeast.

2.3 Variability in the Sea Ice Record

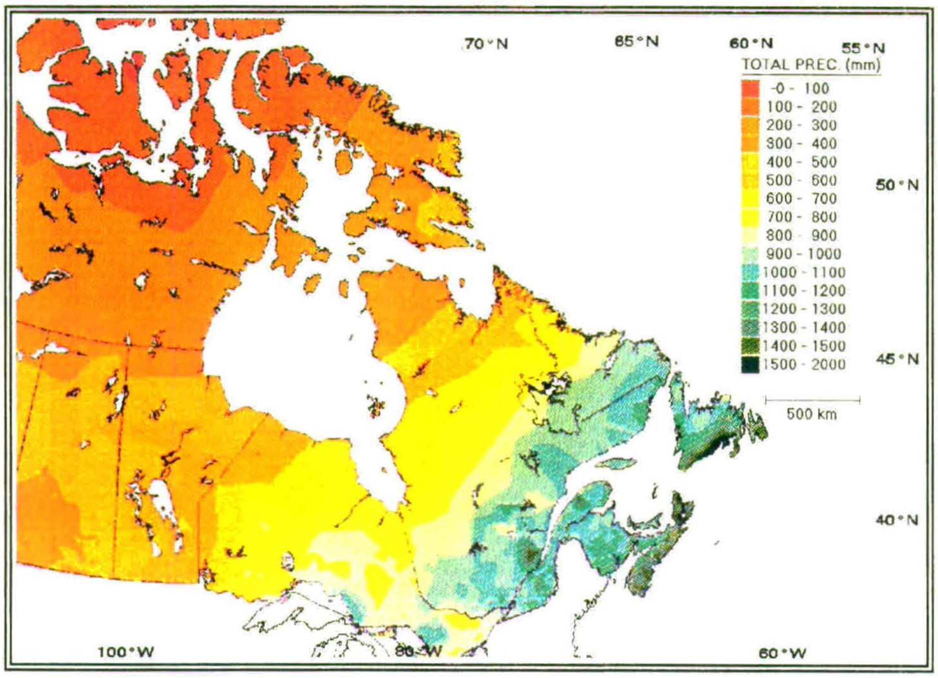
The variability in the temperature record is reflected in the timing of freezeup and breakup dates. *Figure 1.4.10* shows the transition from ice covered to open water conditions for five locations in the Bay (northwest, southwest, southeast, northeast, and James Bay) for the years 1972 through to 1988. There is more variability in the timing of breakup in the spring than freezeup in the fall. The year that had the shortest open water season in most parts of the Bay was 1972. The year that had one of the longest was 1981.

2.4 Comparison between 1972 and 1981

Variability in climate from year to year is of course reflected in all the parameters of the climate system, i.e., the upper air circulation, storm tracks and frequency, air temperature, and sea ice and lake ice. As mentioned in section 1.4.1, the warm 1981 year and the cold 1972 year were the two most extreme years for the region. The differences in the mean position of the trough and in the position of the 510 decametre height contour at the 50 kPa level is shown in *Figure 2.4.1*. The normal position of the trough is just east of Hudson Bay, however, during the winter of 1972 the mean position moved to a line running from Ungava Bay south to the Gulf of St. Lawrence. This eastward displacement of the trough put the region on the west side of the trough resulting in increased frequency of anticyclones and cold northerly flow. In 1981, the trough position shifted considerably west to lie in a north-south line through the centre of Hudson Bay. This westward displacement of the trough allowed increased frequency of southerly flow and increased storm activity especially on the eastern side of Hudson Bay.

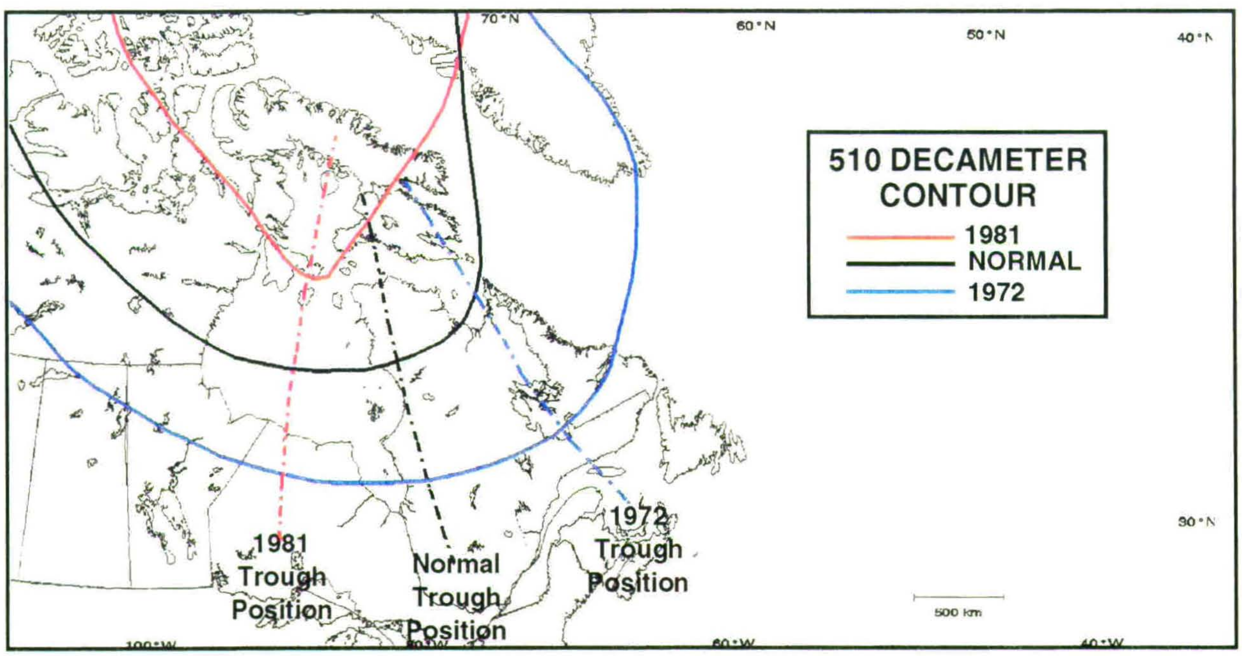
The differences in lower tropospheric temperature are reflected in the differences in the heights of the 50 kPa surface. The much larger area of heights less than 510 decametres in 1972 reflects colder tropospheric air over the entire region. In 1981, the 510 decametre contour only came as far south as Foxe Basin. These lower troposphere temperatures are reflected

Figure 2.2.3 Annual average total precipitation, 1951-80.



Source: Environment Canada

Figure 2.4.1 Position of the winter 510 decameter contour and 50 kPa trough position.



Source: Environment Canada

in the surface temperature anomaly illustrated in Figures 2.4.2 and 2.4.3. The changes in the timing of freezeup and breakup dates and the length of the open water season have already been mentioned in section 1.3.

2.5 Trends

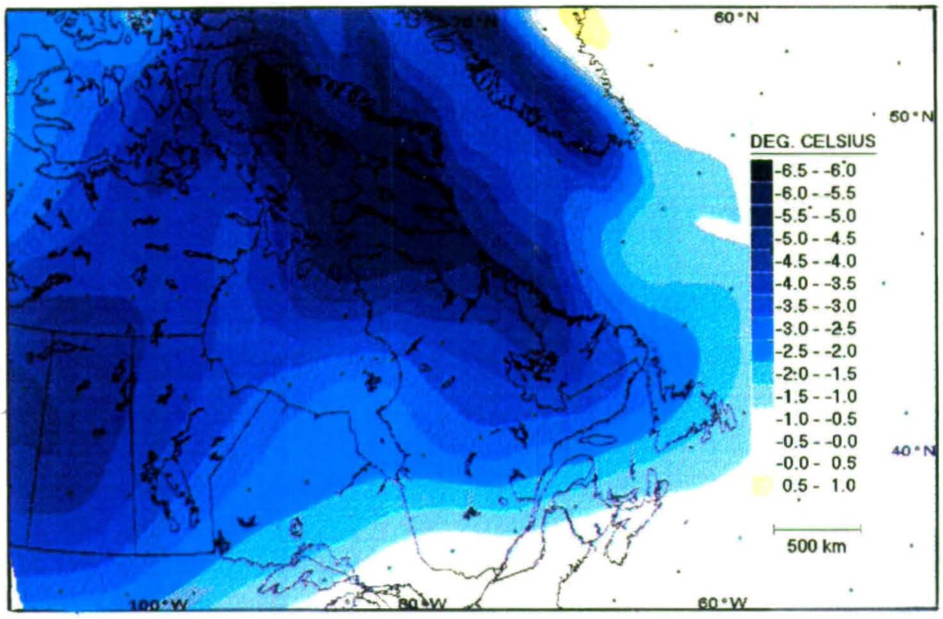
Analyses of world temperature records collected over the past century show that the climate of the Earth is changing. Globally averaged temperature, the most commonly used indicator of climate change, has increased 0.3° to 0.6°C (Houghton et al. 1990). For Canada, mean annual surface air temperatures rose 1.1°C over the same period (Gullett and Skinner 1992). All regions of the country, with the exception of portions of the eastern Arctic Archipelago and the Atlantic coast have experienced the warming, with the highest value being 1.7°C in the District of Mackenzie and the lowest near 0°C in the eastern coastal regions. This observed warming, both in Canada and globally, is of the same magnitude as natural climate variability over the century time period. The observed changes, therefore, could, in whole or in part, be due to this natural variability.

Many factors, both natural and human in origin, influence the climate of the Earth. The Sun provides the driving energy for the Earth's weather and climate. The Earth's surface intercepts solar radiation, including that in the visible part of the spectrum. About one-third is reflected by all of Earth's various surfaces, while the remainder is absorbed by the atmosphere, land, ice, water, and biota. The atmospheric circulation and ocean currents then redistribute this energy around the globe. The energy absorbed from solar radiation is balanced in the longer term by outgoing radiation from the Earth and atmosphere. Otherwise, the Earth would warm up or cool as a result of an imbalance in these quantities. This outgoing radiation takes the form of invisible infrared energy.

The Earth's climate will be affected by any factor which alters the short-wave radiation received from the Sun, the heat radiation released to space, or the redistribution of energy within the atmosphere, and between the atmosphere, land, water, ice, and biota. A number of natural factors can change the balance between the energy absorbed by the Earth and that emitted by it in the form of infrared radiation. Various natural factors lead to climate variability and change, from the short term, for example, volcanism, to the extremely long term, for example, continental drift.

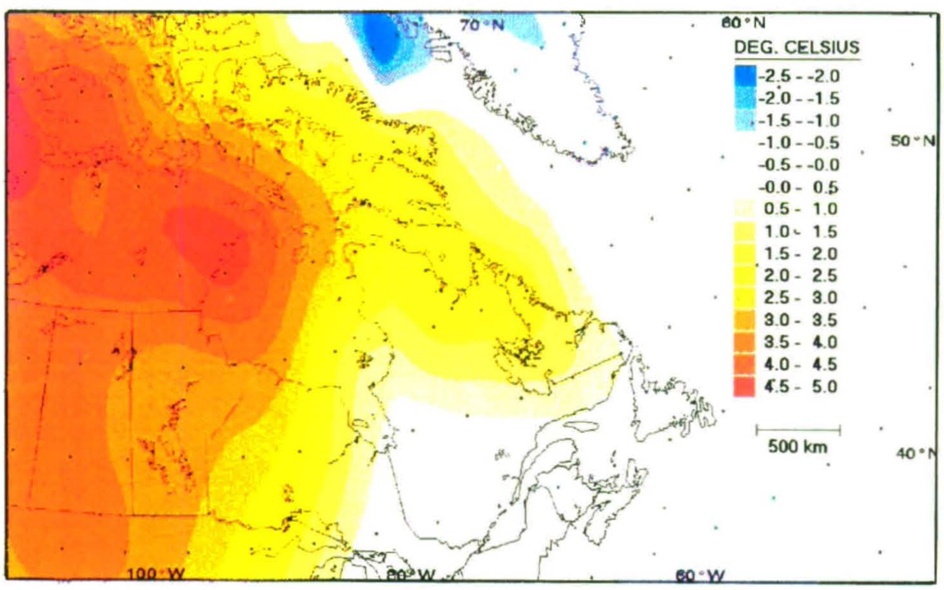
For the past few decades, world observatories have been documenting the increases of certain gases in the atmosphere which

Figure 2.4.2 Winter surface temperature departures from the 1951-80 average for 1972.



Source: Climate Research Unit, University of East Anglia.

Figure 2.4.3 Winter surface temperature departures from the 1951-80 average for 1981.



Source: Climate Research Unit, University of East Anglia.

are known to be extremely effective radiators of energy. These gases have been shown by calculation and climate models to be capable of slowing the transfer of heat energy out of the lower atmosphere, causing a rise in air temperature at the surface, and forming an enhancement to the Earth's natural greenhouse effect.

There are a number of data types considered useful for monitoring climate variability and change. Primary elements such as air temperature and precipitation are the most common. There are also a number of derived elements and combinations of elements. Derived climate data include freezing, thawing, and growing-degree days, frost-free and growing seasons, drought duration, extended hot and cold spells, the humidex, wind chill, and storm frequencies and intensities. In addition to being revealing climate change indicators, these data have also traditionally provided assistance in planning a variety of human affairs. There also exists a wide range of historical documentation of climate-related phenomena, including records of droughts, floods, harvests and other phenological information, glacier extent, sea, river, and lake ice conditions, and weather diaries and ships' logs. Such information can provide an indication of the longer term variability of a climate element.

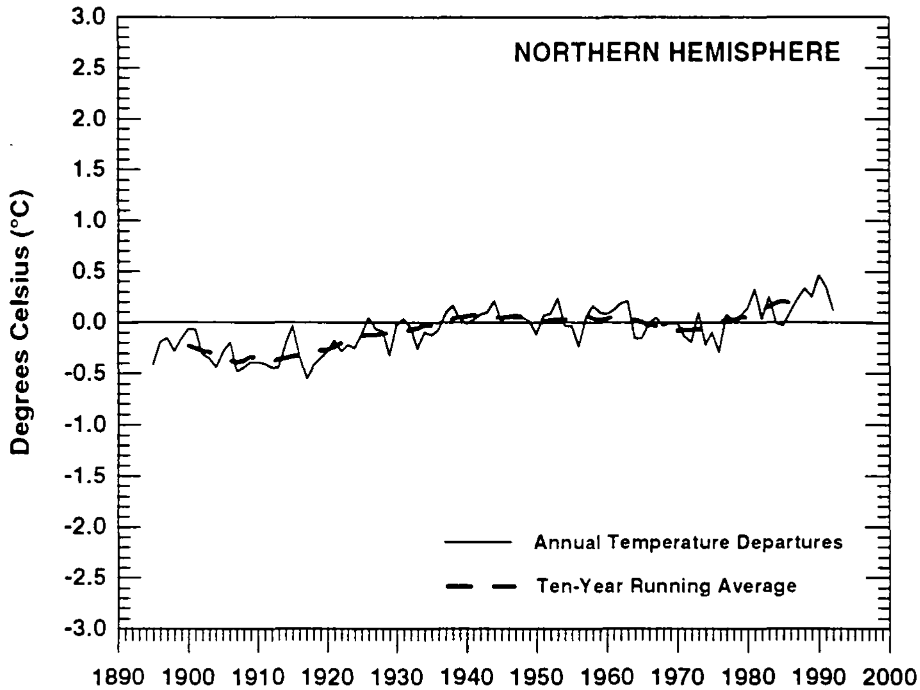
In this report, time series of the observed elements of air temperature and precipitation in the Hudson Bay Bioregion are examined for evidence of change in the recent past. Also, time series of lake ice conditions are examined as corroborating evidence of changing climate conditions during the transition seasons during the recent instrumental period in the region.

2.5.1 Air Temperature

Temperature analyses for the Northern Hemisphere and Canada show similar trend patterns since 1895. Hemispherically, average temperatures have increased about 0.5°C over the past century, while for Canada they have gone up about 1.0°C. The smaller net increase and the lower year to year variability in the hemispheric data are a function of the non-uniform nature of warming over the surface of the globe, and the fact that many more data points are used in the hemispheric averaging. The trends, as indicated by the ten-year running averages, show rising temperatures from the early 1900s to the 1940s, followed by cooling to the 1970s, and then by warming through the 1980s and into the 1990s, are shown for the Northern Hemisphere in Figure 2.5.1, and are mirrored on a much smaller scale, for Canada as a whole, in Figure 2.5.2.

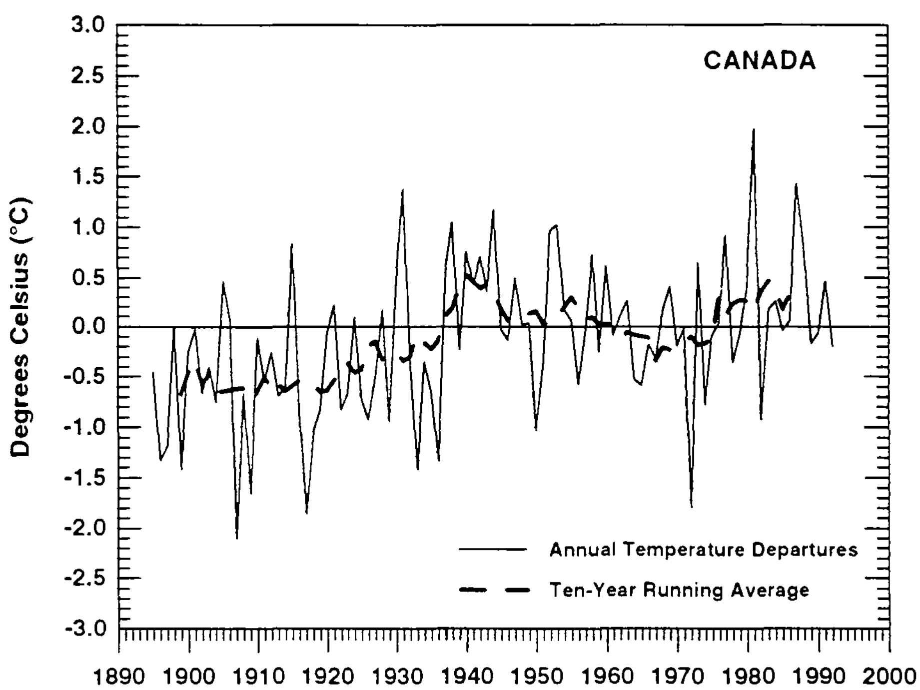
Over the past thirty years, this observed overall warming of the Northern Hemisphere has not been spatially uniform. Figure 2.5.3 shows the trends in mean annual temperature in degrees Celsius per decade for the period 1961 to 1990. Large areas of the Northern Hemisphere landmass have experienced warming.

Figure 2.5.1. Annual mean daily temperature departures from the 1951-80 average.



Source: Climate Researcg Unit, University of East Anglia

Figure 2.5.2. Annual mean daily temperature departures from the 1951-80 average.



Source: Environment Canada

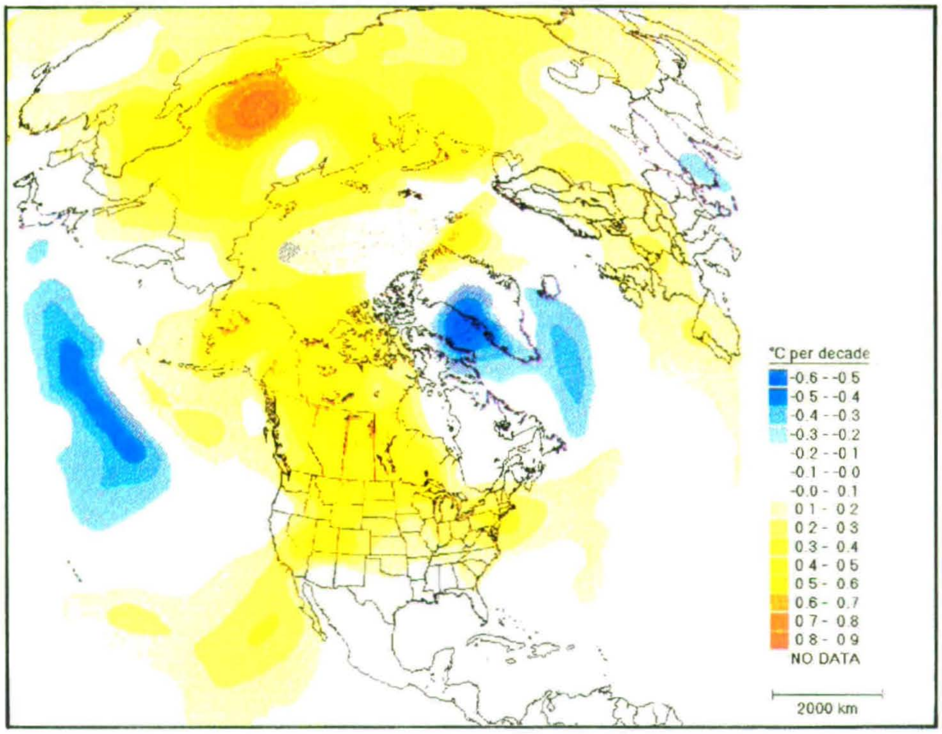
However, large sectors of the North Atlantic and North Pacific Oceans have simultaneously experienced cooling. The Hudson Bay Bioregion is located in a transition area, between an area of anomalously warm conditions (northwestern North America) and an area of colder than normal conditions (northern Atlantic Ocean). Maps of seasonal mean air temperature trends from 1961 to 1990 are shown in Appendix 2.5.1 to 2.5.4. The warming of the Northern Hemisphere landmass has occurred primarily during the winter and spring seasons, with actual large-scale cooling having taken place during the autumn season. The cooling over the North Atlantic and North Pacific Oceans, however, is evident during all seasons.

As previously stated, recent studies have shown that for a large portion of the Northern Hemisphere landmass, including Canada, night-time minimum temperatures are increasing more than are the day-time maximum values during the past 100 years (Karl et al. 1991; Skinner and Gullett 1993). For Canada, annual mean temperatures over the past 95 years have risen about 1.1°C. There has been a statistically significant increase in mostly-nighttime minimum temperatures of about 1.5°C per century and a lower but still statistically significant increase in the mostly-daytime maximum temperatures of about 0.7°C per century. Nationally, seasonal minimum temperatures also show significant increases during all seasons, while coincident maximum temperatures show a significant increase only during the summer. Therefore, maximum and minimum air temperature time series for the Hudson Bay Bioregion are examined in this report for regional evidence of change.

Figure 2.5.4 shows the locations of the fourteen stations used for analysis of maximum and minimum air temperature trends as well as total precipitation trends. The temperature data were extracted from the Historical Canadian Climate Database (HCCD). The HCCD was constructed from the National Climate Data Archive (NCDA) of the Atmospheric Environment Service (AES), utilizing climate stations that were selected on the basis of spatial distribution, length of record, data continuity, homogeneity assessments, and other factors. The HCCD was assembled to provide climate researchers access to an initial but expanding dataset that has been rigorously quality controlled, assessed for homogeneity (Gullett et al. 1991), and adjusted, where necessary, to ensure regional representativeness. The data adjustments that were carried out had the effect of filtering out some of the "local" noise, thereby making the data suitable for use in regional scale analyses. Departures of annual and seasonal values from the 1951 to 1980 normal were calculated for the fourteen locations in the Hudson Bay Bioregion, and then averaged, to create regional series of departures from normal.

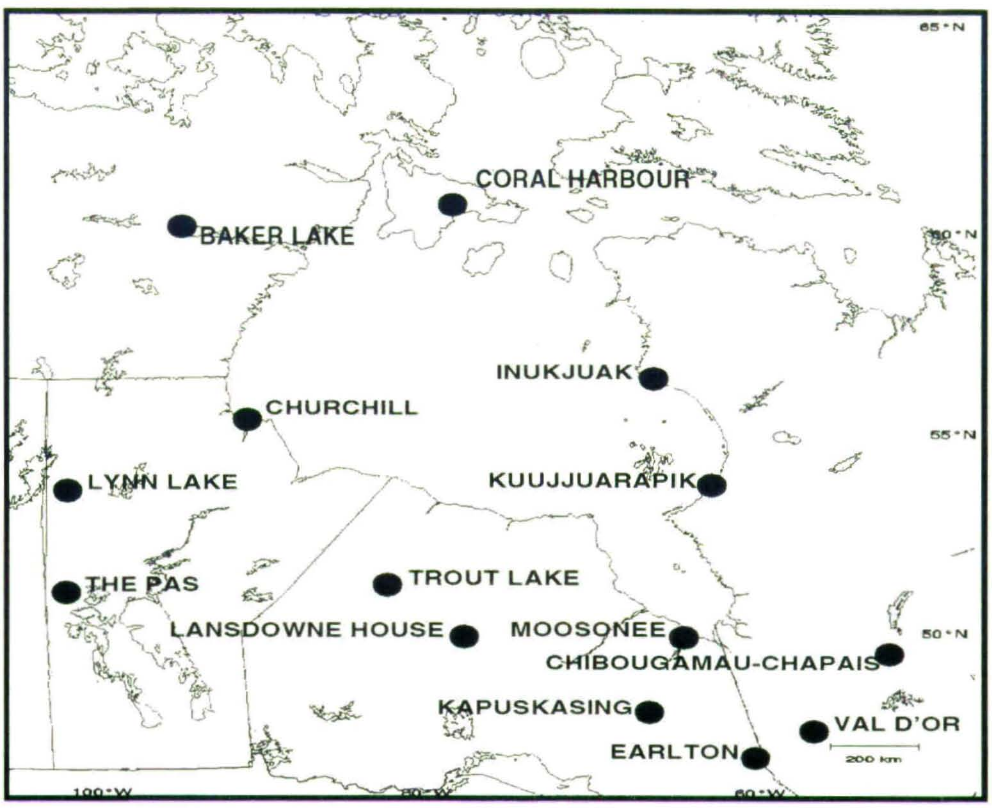
The Hudson Bay Bioregion is poorly represented prior to 1912, and therefore the regional series had to be restricted to the

Figure 2.5.3 Annual surface temperature trends, 1961 to 1990.



Source: Climate Research Unit, University of East Anglia.

Figure 2.5.4 Location of stations used for analysis of regional temperature trends.



Source: Environment Canada

1912 to 1992 period, when three stations provide relatively complete data. Five stations have complete records from 1915, seven by 1926, with all fourteen by 1946.

Figures 2.5.5 and 2.5.6 show the time series of annual daily maximum and minimum temperature departures, respectively, from the 1951 to 1980 average for the period 1912 to 1992 for the Hudson Bay Bioregion. These are accompanied by ten-year running averages to denote trends. No pronounced overall warming trends are evident in either time series such as that indicated in the Canada national time series in Figure 2.5.2. However, both records do reflect the national pattern of warming into the 1940s, followed by cooling into the 1970s, and a resumption of warming through the 1980s. Seasonal time series of daily maximum and minimum temperature departures are shown in Appendix 2.5.5 and 2.5.6. The national pattern is reflected in the winter and spring series with the most pronounced warming since the 1970s being most evident in both the maximum and minimum series during the spring season. No trends are evident during the summer season. Substantial autumn cooling is clearly evident since 1950 in both maximum and minimum temperatures.

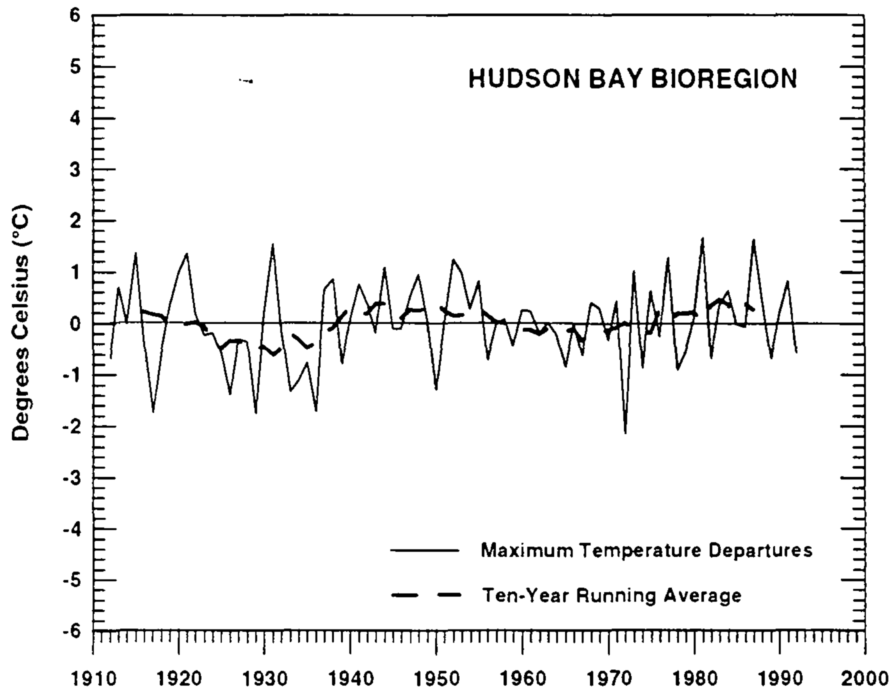
Guiot (1985) shows temperature anomalies and sea-level pressure fields during the instrumental record which correspond with anomalies from the 18th and 19th century, obtained from the proxy record (historical documents). 1972 and 1978 were cold years, and circulation featured low pressure (101.0 kPa) over Baffin and high pressure (101.7 kPa) over Keewatin. During 1953 and 1968, which were warm years, there was a weak meridional pressure gradient (101.3 - 101.6 kPa). Spring 1977 was warm, especially in the southwest (>3C° above average) and there was relatively high pressure throughout the region (101.4 - 101.7 kPa).

Another source of proxy data is wildlife breeding, migration, and feeding patterns. For example, monitoring of marine birds during 1981-1993 at Coats Island, at the north end of Hudson Bay, has not revealed any clear trends in breeding dates. Since 1991, however, part of the diet of chicks has shifted from sculpin to capelin. What these changes might relate to is not clear, but capelin is at the edge of its range in northern Hudson Bay, so its abundance may be a sensitive indicator of marine conditions (Gaston 1993).

2.5.2 Total Precipitation

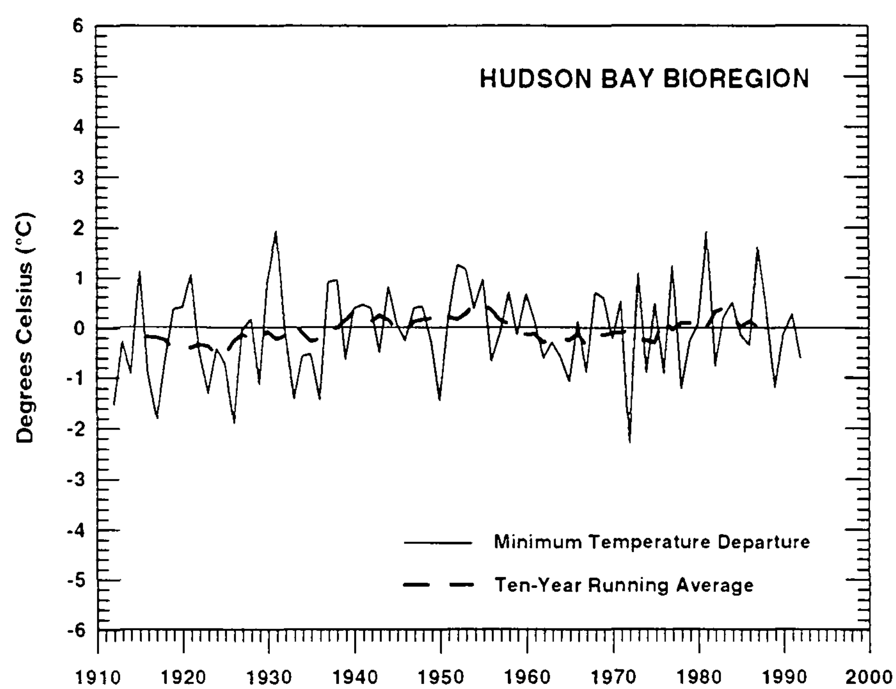
The precipitation data were also extracted from the HCCD. For the Hudson Bay Bioregion, the locations are the same as those chosen to study temperature change. Datasets have been assembled and investigated for completeness, length of record, and data quality. Data sets of obviously poor quality were replaced with better quality values from a nearby site, or were shortened to eliminate periods of poor data. Because

Figure 2.5.5. Annual daily maximum temperature departures from the 1951-80 average.



Source: Environment Canada

Figure 2.5.6. Annual daily minimum temperature departures from the 1951-80 average.



Source: Environment Canada

precipitation can vary greatly over both time and space, the processes of replacing missing values and filtering out non-climatic signals are very complex, and difficult to perform with reliable results. For the current analysis, none of the basic data have been altered in any way. Departures of annual and seasonal values from the 1951 to 1980 normal were calculated for the fourteen locations in the Hudson Bay Bioregion, and then averaged, to create regional series of departures from normal. For precipitation, the Hudson Bay Bioregion is poorly represented prior to 1940, and therefore the regional series had to be restricted to the 1940 to 1992 period, when 90% or more of the stations provide relatively complete data.

Figure 2.5.7 shows the time series of annual total precipitation departures from the 1951 to 1980 average, accompanied by a ten-year running average, for the period 1940 to 1992 for the Hudson Bay Bioregion. This time series shows an overall increasing trend in annual precipitation. The trend indicates generally below normal precipitation prior to the mid-1960s and then generally above normal precipitation into the 1980s, with a slight decline after the mid-1980s. The wettest year was 1979, with over 80 mm more precipitation than normal. The driest year was 1944, with almost 100 mm less than normal. Seasonal time series of total precipitation departures are shown in Appendix 2.5.7. The national pattern towards higher totals is reflected in the spring, summer, and autumn series. However, drier-than-normal conditions prevailed during the 1980s only during the summer season. The winter series shows normal precipitation until the 1970s followed by a gradual reduction in totals through the 1980s.

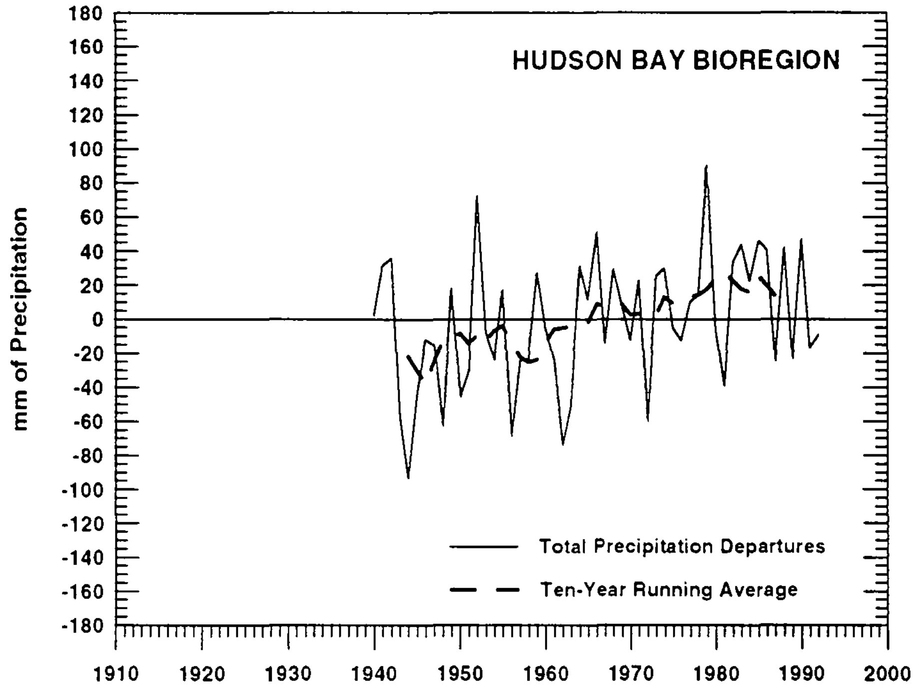
Analysis of global precipitation change (Eischeid et al.1991) indicates that, globally, decades since 1950 have tended to be wetter than those in the first half of the century. As well, there is some indication that annual precipitation appears to have declined in the 1980s over land areas.

For Canada, the effect of the bias introduced by the improvements to the precipitation instruments and measurement techniques is expected to be of some consequence. The magnitudes of these impacts on national and regional data have not been established at this time. It must be emphasized that conclusions based on these preliminary analyses must be tentative at this time. However, the trend towards increasing precipitation in the Hudson Bay Bioregion also appears to be consistent with those identified in many other regions of Canada (Findlay et al. in preparation).

2.5.3 Lake Ice

Five lakes in the Hudson Bay Bioregion with high quality records were selected to show the changes that have occurred in their ice conditions over the years. These lake ice sites are

Figure 2.5.7. Annual total precipitation departures from the 1951-80 average.



Source: Environment Canada

all located in the southwestern areas of the region, as can be seen in Figure 2.5.8. Unfortunately, there are no lake ice observation sites on the eastern side of Hudson Bay. Figure 2.5.9 shows the time series, accompanied by ten-year running averages, for complete freeze over dates for Big Trout Lake, at Trout Lake, Ontario. No trend towards earlier or later freezeup is evident.

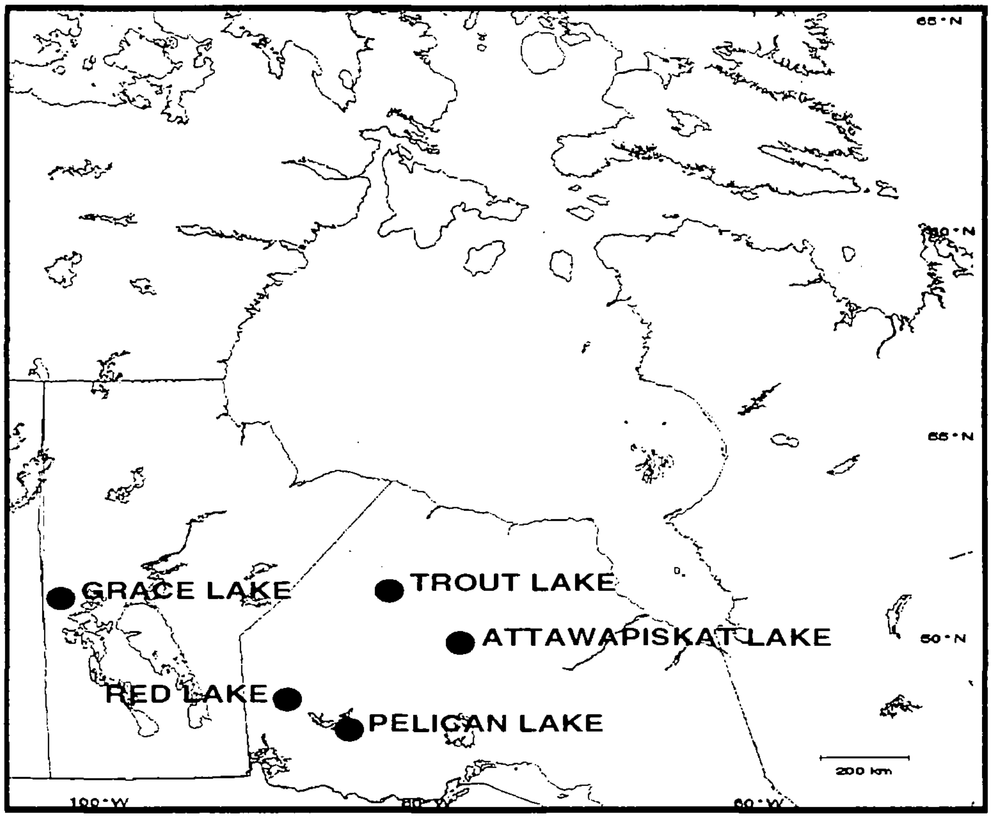
The water clear of ice dates for Big Trout Lake show more pronounced changes, as can be seen in Figure 2.5.10. Breakup dates are occurring about one week earlier, according to the running average trend. Time series of freezeup and breakup dates for the other four lakes are shown in Appendix 2.5.8 to 2.5.9. Each displays the same absence of trend in freezeup dates and general trend towards earlier breakup of ice cover.

The observed recent winter and spring warming appears to have affected the amount of ice forming on the lakes and also the onset of breakup in the spring. The changes occurring in the breakup dates suggest that temperatures associated with the spring and winter seasons are occurring earlier, and are perhaps warmer, than they were in the past in this area of the Hudson Bay Bioregion. The winter and spring season mean air temperature trends from 1961 to 1990, shown in Appendix 2.5.1 and 2.5.2, support these observations. The areas occupied by these lakes have warmed during the winter and spring seasons, while there has been actual cooling in this area during the autumn season, as seen in Appendix 2.5.4.

Schindler et al. (1990) examined climatic and hydrologic records for the past 20 years for the Experimental Lakes Area of northwestern Ontario and found that air and lake temperatures had increased by about 2°C over the period. Also, the ice-free season duration has increased by about 20 days, due mainly to earlier breakup dates in the spring. Fall freezeup dates were not observed to change significantly. Earlier spring breakup of the lakes in this area was seen as the result of two factors, the increased April-May air temperatures and reduced snow cover and warmer temperatures in March causing earlier snow melt and increased solar radiation absorption by the lake in early spring.

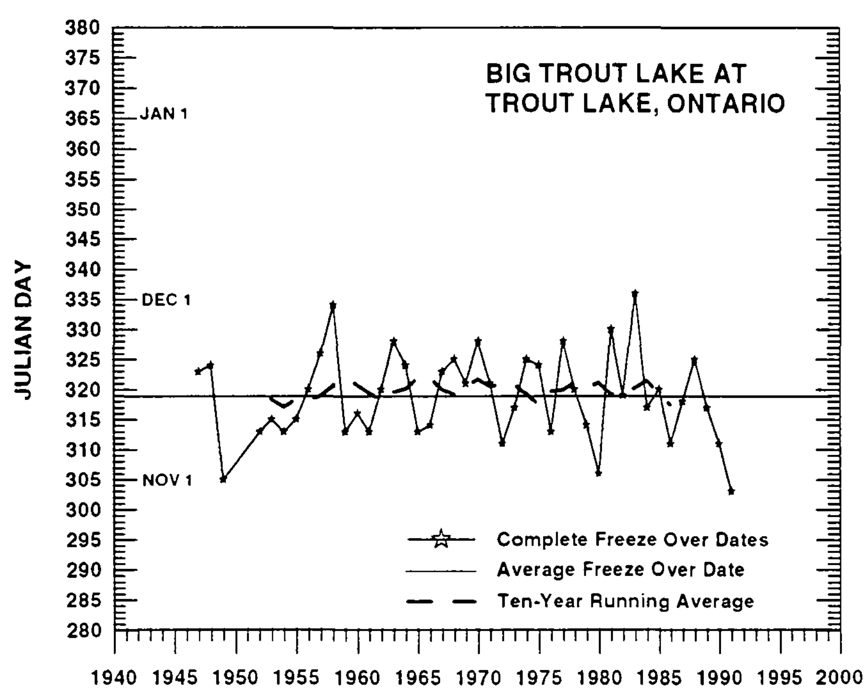
Satellite imagery has recently proven to be beneficial in lake ice studies. Data obtained in this manner provide near-real-time information on the freezeup and breakup of the entire surface of a lake. As more and more meteorological stations become automated, the freezeup and breakup dates cease to be recorded as technicians are no longer present to observe ice conditions.

Figure 2.5.8 Location of lake ice observations.



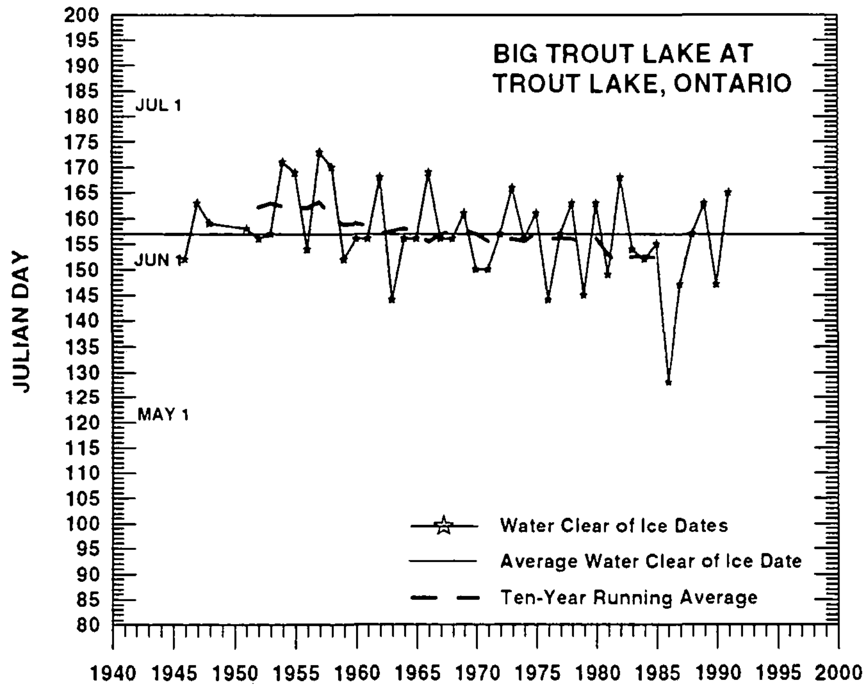
Source: Environment Canada

Figure 2.5.9. Complete freeze over dates at Big Trout Lake, Ontario.



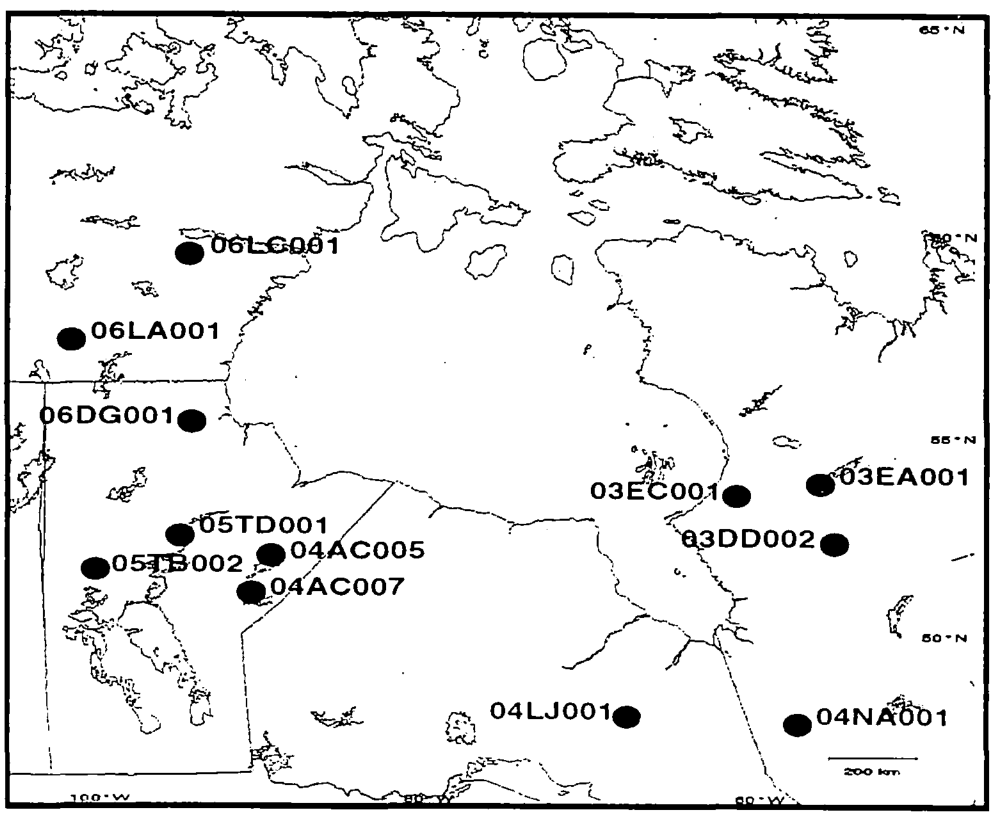
Source: Environment Canada

Figure 2.5.10. Water clear of ice dates at Big Trout Lake, Ontario.



Source: Environment Canada

Figure 2.6.1 Location of long term streamflow stations.



Source: Environment Canada

2.6 Variability in the Streamflow Record

The streamflow at the outlet of a basin integrates the effects of the precipitation and evaporation over the basin area and responds to changes in the regional climate. To determine if there are any trends or patterns in the surface water characteristics of the bioregion which may be attributable to climate change, twelve long term hydrometric stations were selected for analysis. These stations shown in *Figure 2.6.1* and listed in *Table 2.6.1* were chosen based on their record length, data quality, natural flow regime and to provide a reasonably good coverage of the bioregion. The need for long record lengths and good quality datasets for trend analysis is obvious. It is also critical that the flow regime of a basin be natural such that any variability found to occur is attributable to hydrologic processes. Rivers which are described by Environment Canada as having a natural record are intended to reflect the fact that there has been no or minimal man-made interventions in the watercourse. The presence of dams or control structures resulting in a significant change in basin storage would not be considered a natural watercourse.

Table 2.6.1 -- Long term hydrometric stations

	ID	PROV	STATION NAME	LATITUDE	LONGITUDE	BASIN AREA (KM²)	START YEAR	NO. OF YRS
1	03DD002	PQ	DE PONTOIS (RIVIERE) EN AMONT DE LA RIVIERE SAKAMI	53:10:03N	074:28:23W	13200	1960	29
2	03EA001	PQ	BALEINE GRANDE RIVIERE DE LAC A LA SORTIE DU LAC BIENVILLE	54:50:57N	073:59:01W	21000	1962	27
3	03EC001	PQ	DENYS (RIVIERE) PRES DE LA GRANDE RIVIERE DE LA BALEINE	55:00:30N	077:03:50W	4660	1960	30
4	04AC005	MB	GODS RIVER BELOW ALLEN RAPIDS	55:01:35N	093:50:10W	25900	1933	53
5	04AC007	MB	ISLAND LAKE RIVER NEAR ISLAND LAKE	54:03:34N	094:39:34W	14000	1933	55
6	04LJ001	ON	MISSINAIBI RIVER AT MATTICE	49:37:00N	083:15:48W	8940	1920	71
7	04NA001	PQ	HARRICANA (RIVIERE) A AMOS	48:36:02N	078:06:34W	3680	1933	58
8	05TB002	MB	GRASS RIVER AT WEKUSKO FALLS	54:47:20N	099:58:10W	3250	1924	34
9	05TD001	MB	GRASS RIVER ABOVE STANDING STONE FALLS	55:44:35N	097:00:00W	15400	1915	30
10	06GD001	MB	SEAL RIVER BELOW GREAT ISLAND	58:53:30N	096:16:31W	48100	1955	33
11	06LA001	NT	KAZAN RIVER AT OUTLET OF ENNADAI LAKE	61:15:13N	100:58:26W	21400	1962	25

12	06LC001	NT	KAZAN RIVER ABOVE KAZAN FALLS	63:39:10N	095:51:19W	70000	1965	24

The mean annual streamflows for the twelve stations were extracted from HYDAT, Environment Canada's historical surface water database. These mean annual flow values were converted to an average depth of water in millimetres over a drainage basin area to allow for easier comparison with precipitation data. Time series of the mean annual flows with a ten year running mean were plotted for each station. These are shown in *Appendix 2.6*. Visual inspection of these time series shows considerable interannual variability, but no significant trends or patterns were found. The two north western stations, 06LA001 and 06LC001, suggest a slight increasing trend but with only 25 and 24 years of data, they were not considered to be reliable. Other studies by Rouse (1993) and Anderson et al. (1992) have applied more sophisticated statistical testing for trends on station 04LJ001, which has the longest record length with 71 years. Neither study found any significant trend in its mean annual flow time series.

Although mean annual flow does not appear to exhibit any long term trends, changes in the seasonal cycle may be occurring. The Missinaibi River is currently experiencing an earlier spring peak (Christie, 1993).

3. CLIMATE SCENARIOS

3.1 Introduction

Most scientists argue that there is a high probability of major global warming during the next century. But they also acknowledge that uncertainties in our understanding of the climate system and its complex behaviour make it difficult to project with confidence how much warming will occur, how quickly it will take place, or what its regional characteristics will be. Resolving or reducing these uncertainties has become a priority for current climate research efforts in Canada and elsewhere.

One of the central features of this research has been the use of numerical models - and, in particular, powerful computer-based representations of the global climate system known as general circulation models (GCMs) - to estimate the effects of increasing greenhouse gas concentrations on the world's climate. These models have added greatly to our knowledge of climate change.

Studies into the Earth's past climates can provide valuable information on the natural variability of the global climate system, and on the response of ecosystems to climate change. However, they cannot answer the many questions about processes and feedbacks within the very intricate climate system, and between its sub-elements. On the other hand, planned experimentation with the Earth's climate is virtually impossible, as well as foolhardy, because of the immense size of the system and the risks associated with such experiments. Although humans through their actions have already inadvertently started an unplanned global experiment with the climate, the results of that experiment will take decades and centuries to emerge. The knowledge gained would be too late for use in developing mitigative and responsive policy action.

Thus climate models have emerged internationally as the only practical and timely manner of investigating the behaviour of the complete climate system and studying its response to the forces that affect it.

3.2 Modelling the Global Climate System

GCMs simulate climate by means of equations representing the physical processes of the climate system. The basic processes are those involving radiation, heat and motion, and the water cycle. The model must also calculate the interactions between these processes. For a more detailed discussion of this topic the reader is referred to a recent publication by Environment Canada (1993).

GCMs must also accommodate the spatial diversity of a large three-dimensional world varying with time. They try to capture as much of this diversity as possible by making calculations for a large number of regularly spaced points, both on the Earth's surface and in the atmosphere. The more points there are, the higher the resolution of the model and the finer the detail it can simulate. Today's highest resolution models may have 30 or more vertical layers, with each spatial point representing a few degrees of latitude and longitude horizontally, or an area of about one hundred thousand square kilometres.

In spite of its sophistication, a GCM is still only an approximation of reality. Even the most powerful supercomputers available today cannot handle all the detail needed to give a complete description of the climate system. Nor do we fully understand all of the processes that affect climate. Therefore, the model must simplify or estimate some features of the system and simply ignore others that are not considered important enough to significantly affect the outcome of the model's calculations.

All of the major features are recognizable, but much of the fine detail is missing. The objective, as the models evolve, is to fill in more of the missing detail and make the models more realistic.

3.3 Canadian Climate Centre GCM (GCM2)

The Atmospheric Environment Service's GCM2 (Environment Canada 1993; Boer et al. 1992; McFarlane et al. 1992; Hengeveld 1991) is a second generation GCM, with higher resolution than the more coarse first generation GCMs, more detailed representation of features and processes, and more sophisticated parameterizations. GCM2 incorporates all of the features considered state-of-the-art for a second-generation model of the atmosphere coupled to a mixed ocean layer model and a thermodynamic sea ice model. It has a relatively high horizontal resolution of about 600 km giving a grid of approximately 3.75° x 3.75° in physical space.

It simulates horizontal heat transport in oceans that provides a better representation of the oceanic heat distribution and its effects on regional climates. Sea ice is also simulated more realistically, in terms of both its optical characteristics and its formation and melting. The treatment of clouds is highly sophisticated, incorporating a process for determining cloud optical properties and varying them in response to other climatic changes. The model also reproduces the cycle of night and day and its effects on temperature, pressure, and other climatic elements.

To account for the effects of different types of soil and vegetative cover on the Earth's radiation balance, the Canadian model includes more than 20 different reflectivity variations for land surfaces. In snow-covered regions, the model also accounts for the effects of vegetation and the age of snow on surface reflectivity. In addition, the detailed representation of soil and vegetation types gives a much more realistic modelling of soil moisture capacity.

The most common application of GCMs is to determine the sensitivity of the climate system to a change in one of its key elements. Strictly speaking, the result is not a prediction of climate change, though it does have some predictive value. Instead, it is an answer to a "what if" question - in the case of global warming, what would be the response of the climate system if the concentration of greenhouse gases in the atmosphere increased significantly?

The classic experiment for testing the climate's sensitivity to higher greenhouse gas concentrations is known as a 2 x CO_2 or doubled carbon dioxide equilibrium response experiment. It is, in effect, two experiments in one. It begins with a present climate simulation. The average global carbon dioxide concentration is set at the present value (1 x CO_2) and the model is run until it produces a stable climate. The model is then reprogrammed, with the carbon dioxide concentration set to twice its present level (2 x CO_2), and run again until it reaches a new equilibrium. The difference between the two sets, or the change from the first to the second set, of results is the model's projection of the climate's equilibrium sensitivity to a doubling of carbon dioxide.

3.4 GCM2 Results and the Hudson Bay Bioregion

Numerous models have been developed and are in use around the world today. All of the models agree on the general nature and direction of change - increased surface warming with greater warming towards the poles, cooling in the stratosphere (the part of the atmosphere about 10-50 km above the Earth), increased precipitation and evaporation, and less sea ice. Most also agree on diminishing soil moisture in northern mid-latitude continents in summer. They disagree significantly, however, about the details. Their projections of average global temperature change vary between 1.7°C and 5.2°C. Most results are close to 4.0°C. Recent studies using more detailed cloud processes give results ranging from 2.0°C to 3.5°C. Projections for average global increase in precipitation (and evaporation) range between 3% and 15%. Disagreement on the geographical distribution of these and other changes is also substantial, particularly in an east-west direction.

When the 2 x CO_2 experiment was run on GCM2 (Environment Canada 1993; Boer et al. 1992; McFarlane et al. 1992; Hengeveld 1991), it showed an average global surface warming of 3.5°C, with the greatest warming occurring at the poles in winter. However, temperatures in the stratosphere became cooler. World cloud cover decreased by 2.2% while precipitation and evaporation increased globally by 3.8%. But with more of the additional rain falling over the oceans rather than the land, soil moisture decreased by 6.6%, indicating a probable rise in the frequency of drought in many parts of the world, including Canada. Sea ice also became thinner and retreated poleward, resulting in a loss of about 66% of the sea ice mass.

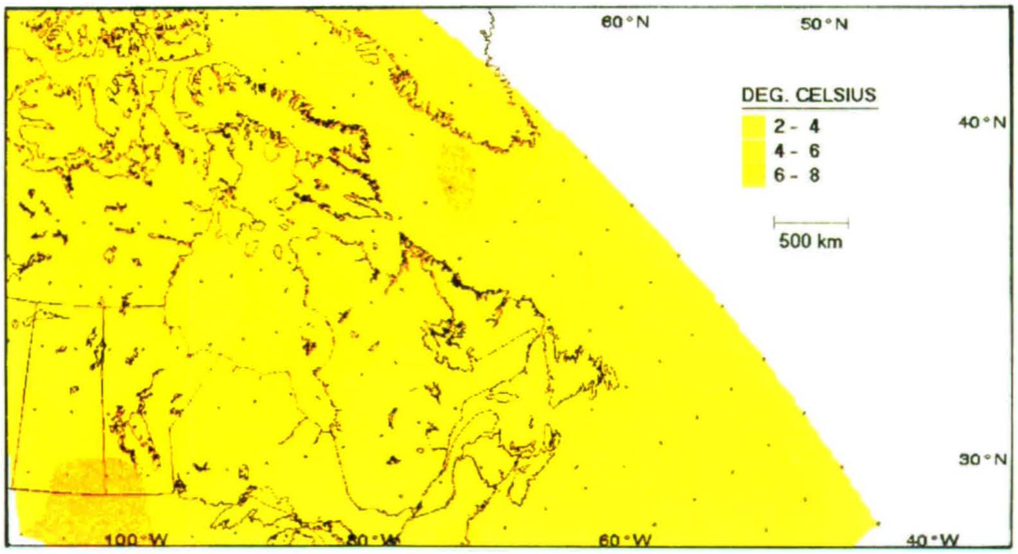
Over Canada the changes were more intense. The model showed daily mean temperatures in southern regions nearly 5°C warmer throughout the year and northern regions as much as 8-12°C warmer in the winter. Although seasonal increases in water supply were indicated for the west coast, the Yukon, and much of the Arctic, there was a decrease of more than 20% in soil moisture for the rich farmlands of the south-central region.

Figures 3.4.1 and 3.4.2 show the annual mean daily maximum and daily minimum temperature surface air temperature changes associated with GCM2 for the Hudson Bay Bioregion. Annual daily maximum (mostly daytime) temperature changes are in the 2-6°C range throughout the region with generally higher increases in the north, western, and northeastern areas, and lower increases in central Quebec and Ontario. Annual daily minimum (mostly nighttime) temperature changes are generally higher, in the 4-8°C range throughout the region. Minimum temperature change is expected to be more uniform over the region with higher increases over central and eastern Hudson Bay. Annual mean daily minimum temperature increases are greater than those of mean daily maximum temperature in areas where the sea ice has diminished or retreated entirely. This is evident in Hudson Bay and also in Baffin Bay between Baffin Island and Greenland. Atmospheric warming is prevented by the presence of sea ice. Daily minimum temperatures are therefore lower in the 1 x CO_2 climate than in the lower ice regime of the 2 x CO_2 climate (Zwiers 1993).

Appendix 3 shows the seasonal mean daily maximum and daily minimum surface air temperature changes associated with GCM2 for the Hudson Bay Bioregion. Changes in minimum temperature exceed those of maximum temperature in all seasons. Large increases of up to 16°C are evident in winter and up to 10°C in spring over Hudson Bay and Baffin Bay. Summer increases are more moderate and generally more uniform between maximum and minimum temperatures with the exception of central Ontario and central Quebec. Autumn temperature increases for both maximum and minimum temperatures are also moderate, from 2-4°C, across the southern areas of the region, and higher, 4-6°C, in northern areas. There is a strong south to north positive gradient in temperature increase during autumn due to a large reduction in arctic sea ice.

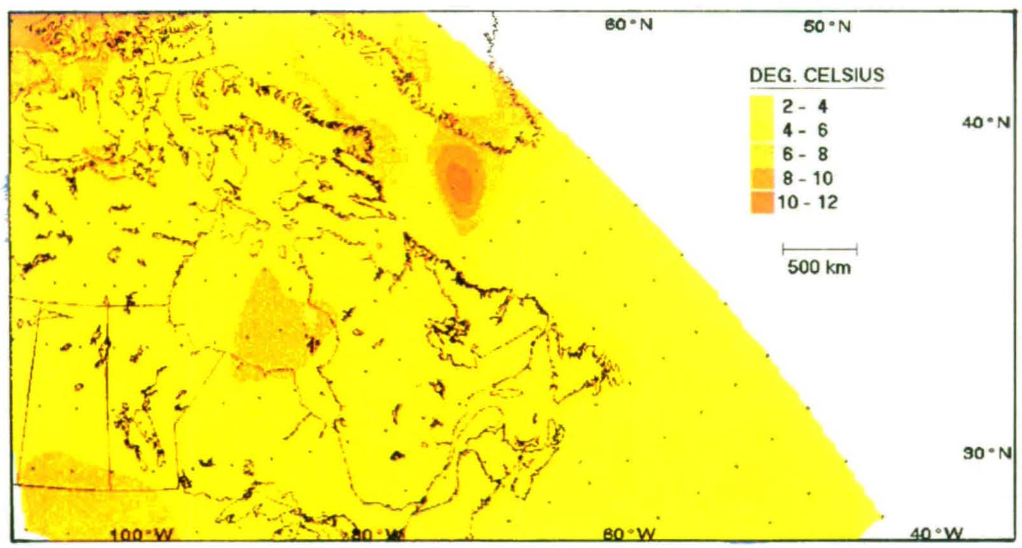
Figure 3.4.3 shows the annual total precipitation changes associated with GCM2 for the Hudson Bay Bioregion. There are moderate increases, 0% to 25%, across the entire region with more substantial increases, 25% to 50%, over eastern Hudson Bay due to less ice cover on Hudson Bay. Appendix 3 shows the seasonal total precipitation changes associated with GCM2 for the Hudson Bay Bioregion. In general, there are modest precipitation increases across the region with some important differences. There are modest decreases in precipitation totals in winter and spring over an area on the northwestern shores of Hudson Bay and during summer in a broad band across the southern areas of the region. Precipitation totals at these times of year are normally quite low, especially in the western areas, and further decreases

Figure 3.4.1 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for annual daily maximum temperature.



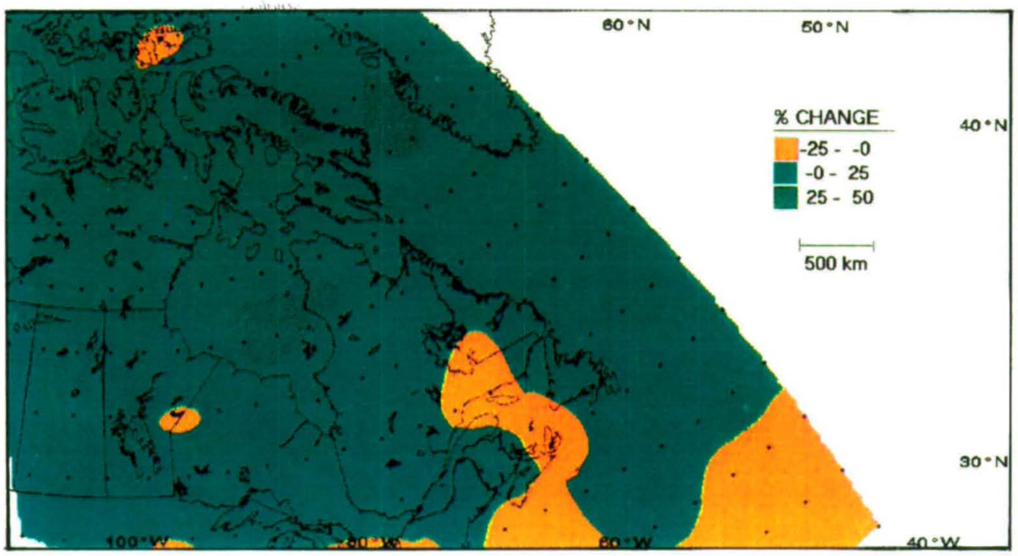
Source: Environment Canada

Figure 3.4.2 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for annual daily minimum temperature.



Source: Environment Canada

Figure 3.4.3 CCC GCM total annual precipitation change from $1xCO_2$ to $2xCO_2$.



Source: Environment Canada

could have significant impact. Autumn increases are substantial, 25% to 50%, over central and western Hudson Bay.

Canadian modellers are now at work on a third-generation GCM. It will incorporate improvements to many of its present features while adding some important new capabilities, such as refining existing parameterizations, particularly the treatment of clouds. Other improvements will include a much more detailed treatment of processes in the middle atmosphere (up to 85 km above the Earth's surface), particularly chemical interactions involving trace gases such as sulphur dioxide, ozone, and carbon dioxide and their effects on the Earth's radiation budget and atmospheric circulation. Also, regional sub-models are being developed to provide a more detailed simulation of local climates.

The most significant feature of the third-generation model, however, will be the inclusion of a three-dimensional, fully circulating, interacting ocean. The new model will allow for the absorption and storage of heat by the deep ocean over periods ranging from a few decades to centuries. This feature will make it possible to conduct a much more realistic type of experiment in which the climate changes continuously in response to gradually increasing levels of greenhouse gases rather than to a sudden leap in concentration as in equilibrium response simulations. This experiment is known as a transient response experiment, and its chief benefit is that it permits the simulation of climate change as we are likely to experience it. These experiments not only should provide a more useful indication of how the climate will evolve but also should help to identify a fingerprint, or the spatial signature of climate change that is indeed taking place.

The benefits of modelling developments are expected to be enormous. The improved capability of GCMs in accurately describing the behaviour of the climate system should help provide more accurate estimates of the rate and regional characteristics of future climate change in Canada, particularly if coupled with high resolution regional climate models.

4. IMPLICATIONS OF CLIMATIC CHANGE

Given the warming scenario illustrated in section 3, we now address the question of regional impacts. Although there have been many first-order impact studies done in other parts of Canada (e.g., water resources, sea ice), few have been attempted in the Hudson Bay area. There have been broad assessments of Arctic impacts recently completed by Woo et al. (1992a,b) and Woo and Gregor (1992), and a case study of the Moose River (Rouse 1993).

Impacts on ice, streamflow, permafrost, and vegetation will likely be linked because of the natural connections that exist in ecosystems. Although climatic change could have direct implications for ice, streamflow, etc., the response of each of these parameters will be influenced by climatic effects on other parameters (vegetation, etc.). This report presents a "list" of impacts, but each should be seen in its ecosystem context.

4.1 Ice Cover

4.1.1 Freshwater Ice

Potential effects of warming would depend on snowfall. In a study of the Mackenzie River, it has been suggested that open water conditions would be extended by 3-6 weeks (Lonergan et al. 1993). Although breakup should occur earlier, heavier spring snowfall would create high albedo conditions in spring, which could counteract the temperature effect (Adams 1992).

4.1.2 Marine Ice

The biosphere is inextricably linked with physical processes which occur over the Bay, especially formation and ablation of sea ice. Sea ice provides physical controls on the productivity, distribution, and interaction amongst various trophic levels. At the simplest level, the open ocean and sub-ice primary production is controlled by the amount of solar radiation that can pass through snow-covered sea ice into the euphotic zone. This control that the physical environment places on the ecosystem of the Bay occurs all the way up the food chain web to the highest trophic levels (i.e., avians, seals, polar bears, and man). At each level the physical environment controls or determines activities. For example, seals often have their lairs in the complicated ridge structures of first-year ice. Polar bears use sea ice as a platform for foraging and hunting. Recurring open water areas in sea ice (polynyas) are biologically favourable areas where high concentrations of wildlife exist. Such areas are usually referred to as arctic "oases" and contribute very much to the biological resources of the Bay.

Changes in the sea ice of the area will have major implications for ecosystems of the Bay. Some GCMs suggest complete disappearance of sea ice in winter! This would eliminate the most important component of the freshwater budget of the Bay (i.e., sea ice melt) and threaten ecosystems as they presently exist. Because very few climate change impact studies for the area have been done, details of potential impacts are not known.

A potentially important feedback on the bioregion could result from earlier melt, since this would lead to a warming of onshore winds. Higher air temperatures during the growing season at near coast and inland locations, accompanied by higher rate of evapotranspiration from the wetlands, could lead to some wetlands becoming completely dry during the longer growing season. This happened during the summer of 1989 at a site 12 km inland. The long term effects of drier wetland would be reduced water yield for streamflow, movement of forest cover northward, and possible increased incidence of fires which would degrade the permafrost (Rouse 1991).

4.2 Permafrost

As the annual mean air temperature increases, the ground temperature near the surface increases, and through conduction, the energy is transferred to the lower levels. The active layer would deepen, with possible disappearance of shallow discontinuous permafrost from the southern margins of the region (Lewkowicz 1992). The warming of the permafrost at the lower (deeper) levels may not occur until long after the ambient air temperature has begun to increase. A scenario of northward retreat of permafrost is illustrated in *Figure 4.2.1*.

In the discontinuous zone, where the permafrost temperature is near freezing, techniques that pay special attention to frozen soils and preservation of permafrost will be required in several engineering activities. They will include foundation design and utility development, highway and railway construction, pipelines construction, the extraction of minerals, and the disposal of waste (Woo et al. 1992a).

With the retraction of the permafrost in the discontinuous zone and increased depth of the active layer in the continuous zone, frost heave will create disruption to the surface. Mud slides and thaw settlements are likely to occur (Gerwick 1990). The frost heaves will impact on almost all aspects of the built environment.

4.3 Streamflow

A 2°C warmer climate would probably lead to reduced flow in the Moose River if annual precipitation was to change by less than +10% (Rouse 1993). The greater the warming, the greater the reduction in flow unless additional precipitation occurs.

The scenario described in section 3 indicates greater warming and higher precipitation than the scenario used by Rouse (1993). It is likely that this will result in higher annual runoff with an earlier spring peak due to earlier snowmelt. However, this scenario has not been rigorously assessed, and further study is warranted. Any changes in streamflow would affect hydroelectric

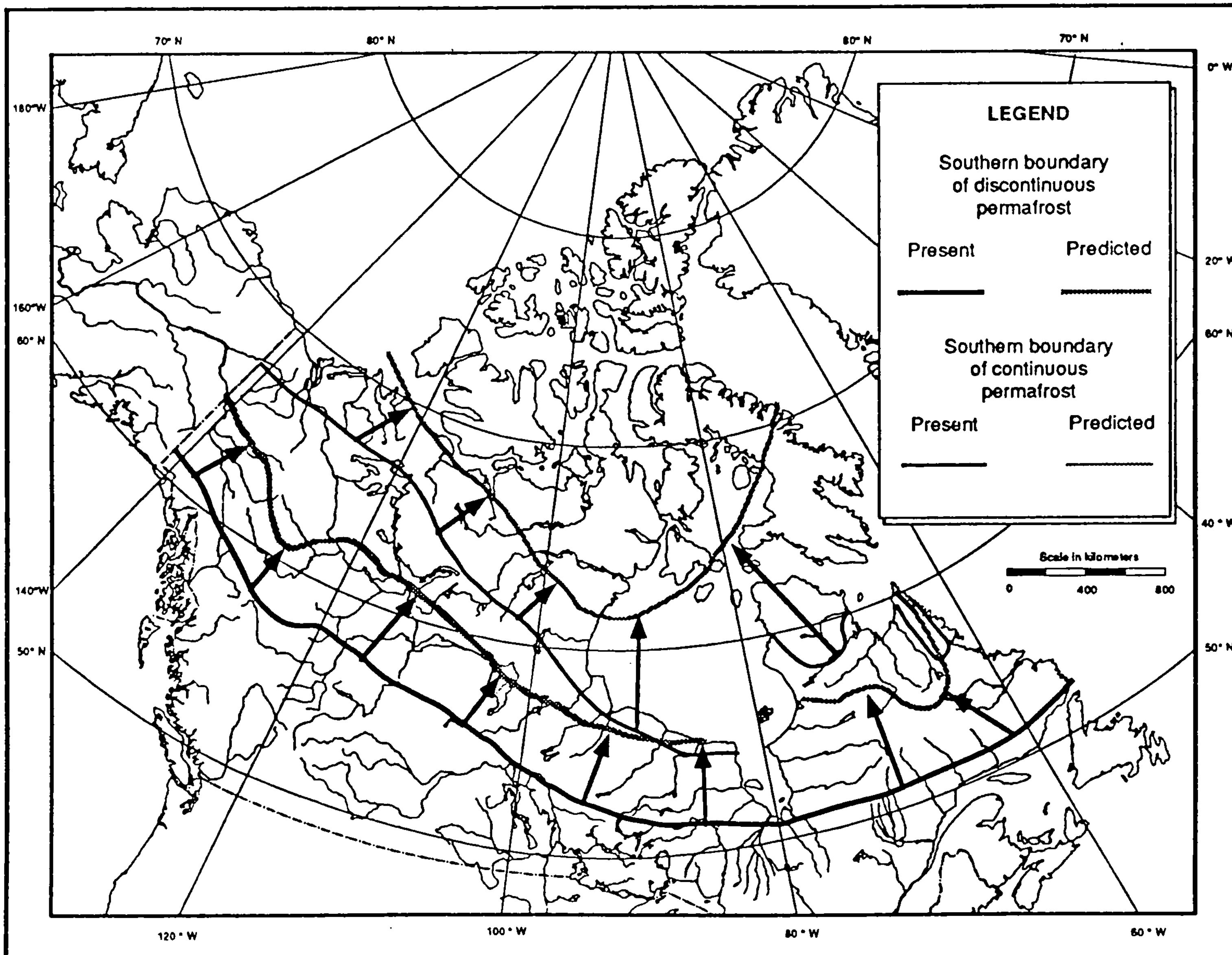


Figure 4.2.1 Shift in the boundaries of discontinuous and continuous permafrost as a result of surface temperature change of 4-5°C (Woo et al., 1992a).

power production and reservoir supply on the Mattagami and Abitibi Rivers, as well as facilities in Quebec and Manitoba.

Changes in runoff would also contribute to changes to the freshwater budget of the Bay. Spring melting of sea ice and snow covering the ice is currently larger than freshwater addition to the Bay from runoff, and the total marine-based melt is largely dependent on the amount and character of ridging over the sea ice (Prinsenberg 1988). The freshwater budget of the Bay could change significantly if warmer temperatures due to changing climate produce thinner sea ice over the winter, thereby altering the current balance between marine-based and land-based inputs.

A different kind of climatic change scenario concerns the potential impacts of the proposed GRAND Canal, which would transform James Bay into a freshwater lake. The reduction in salinity and disruption of coastal currents could result in a delayed ice melt in spring, leading to cooler, wetter summers in the region. Over time, this could lead to a retreat of forests from the James Bay coast, and growth in permafrost (Rouse et al. 1992).

4.4 Vegetation

Black spruce has exhibited increased height during the 20th century, but also increased basal abrasion, suggesting that 20th century warming has been accompanied by increased windblown snow. Older stunted spruces did not exhibit increased growth rates this century because they could not capture enough drifting snow (Lavoie and Payette 1992).

Other paleoecological records indicate that the tree line responded in an asynchronous manner to climate change, with northwestern Canada showing an earlier northward migration and southern retreat of trees than exhibited in the Hudson Bay region (MacDonald and Gajewski 1992). A warming would lead to a delayed response, with coniferous trees migrating northward, with a corresponding reduction in shrub and tundra (Edlund 1992), but an increase in fire could lead to invasion by deciduous trees especially along northward flowing rivers, as has been documented in the Mackenzie Delta (Landhäusser and Wein 1993).

Paleoecological evidence from the treeline northeast of Great Slave Lake suggests that changes in vegetation and lake characteristics can be rapid (decades) if the future climate warms at a faster rate than previous episodes of warming 4-6000 years ago (MacDonald et al. 1993). Using a climatologically-based ecological classification system and a GCM-based scenario of climatic warming, Rizzo and Wiken (1992) suggest that the boreal ecoclimatic province would become discontinuous, and be replaced along the western shore of Hudson Bay by temperate and grassland ecoclimatic provinces. Associated changes in vegetation would occur more slowly, depending on soil conditions.

4.5 Wildlife

The presence of sea ice provides the opportunity for polar bears to successfully hunt seals for food. Reduction in sea ice in Hudson Bay would increase nutritional stress on polar bears, and would force them to migrate northward where the open water season would still be relatively short (Stirling and Derocher 1993).

5. CONCLUSIONS AND RESEARCH NEEDS

This review of climate and climate impacts has been hampered by a sparse monitoring network and incomplete or missing data records. These influence our main conclusions about the Hudson Bay Bioregion's climate, which are as follows:

a) *CURRENT REGIONAL CLIMATE:* The region's cold climate is largely a result of its high latitude continental location, far removed from the Pacific Ocean and other sources of milder air. The Bay plays only a secondary role in determining the position of the upper air trough which controls atmospheric circulation in this region. The Bay does exert some influence at the regional scale, particularly in generating elevated precipitation in downwind areas (primarily the Quebec shoreline) and temperature reductions in the coastal zone during summer due to the presence of sea ice as late as July.

b) *CLIMATIC INFLUENCES ON REGIONAL LANDSCAPES:* The cold climate has led to the development of discontinuous and continuous permafrost everywhere except south of James Bay. This, along with high snowmelt and the presence of ice in lakes and rivers, influences the regional hydrologic regime, which encourages large spring peak flows and the development of wetlands in lowland areas.

c) *SEA ICE:* Sea ice is the largest component of the freshwater budget of Hudson Bay. It also provides the physical controls on biological productivity, distribution, and interaction for all levels in the food web.

d) *SHORT TERM VARIATIONS AND LONG TERM TRENDS:* This region is located in a transition zone between an area of anomalously warm conditions (northwestern North America) and an area of colder than normal conditions (northern Atlantic Ocean). Certain regional trends are evident, particularly the warming in spring; however, these trends are of a lesser magnitude than current variability, as illustrated by the analysis of 1972 vs. 1981. There is no clear regional precipitation trend, but data analysis is affected by snowfall undercatch and changes in instrumentation. Some lakes in the south and southwest are exhibiting earlier breakup of ice, and an earlier spring peak is being observed at the Missinaibi River. Recent cooling has occurred in the Baffin Island region, but this zone is too far east to be noticeable in the Hudson Bay region.

e) *SCENARIOS OF CLIMATIC CHANGE:* The Canadian Climate Centre's General Circulation Model (CCC-GCM2) simulation of CO_2-induced climatic change shows higher temperatures in all seasons, accompanied by the disappearance of sea ice from the Bay. There are uncertainties associated with current GCMs, particularly the lack of a fully circulating interacting ocean and poor treatment of clouds. Efforts are under way to overcome these deficiencies. Although surrogate (proxy) data have provided valuable information on past trends, and can be used to generate 'analogue' scenarios, they do not provide information on CO_2-induced climatic changes induced by human activities. GCMs represent the best source for these scenarios.

f) *IMPLICATIONS OF SCENARIOS OF CLIMATIC CHANGE:* Changes to the hydrologic regime, either from increased hydroelectric development or a CO_2-induced warming of the climate, could affect freshwater and sea ice, discontinuous permafrost, wetlands, vegetation, and the coastal zone. Sea ice could also be directly affected by warmer temperatures. Current predictions from GCMs are that the amount of sea ice that forms in winter would be drastically reduced and may perhaps even disappear under a doubled CO_2 concentration. These changes could have major implications for ecosystems of the Bay, but few studies have been completed to date. What can be said is that there are climate-sensitive areas that could be affected by relatively small increases in temperature, particularly i) permafrost warmer than -2°C, ii) coastal areas up to 10 km inland, iii) wildlife dependent on sea ice and wetlands, iv) infrastructure built in the discontinuous permafrost zone, and v) communities dependent on wildlife. It should also be noted that scenarios of climate warming, hydroelectric development and the GRAND Canal, would probably lead to very different impacts on the bioregion. For example, warming might lead to a reduction in sea ice, but the GRAND Canal's freshwater impoundment, with its reductions in salinity, could have the opposite effect.

We conclude with some recommendations:

1) There should be increased monitoring of snow cover, streamflow in unregulated rivers, freshwater ice, the Bay's freshwater budget (including snow cover and snow melt over the sea ice), and ground temperatures, so that significant changes

in permafrost, hydrology, and salinity could be detected. This would complement ongoing efforts at monitoring temperature, sea ice and other parameters. These efforts should be protected and maintained over the long term.

2) An integrated regional assessment of scenarios of climatic changes and water resources development should be carried out. This could focus on climate-sensitive areas and their various stakeholders (governments, communities, resource industries, ecosystem maintenance, transportation, tourism).

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APPENDIX 1. CLIMATE SUMMARIES

A1.5 Ground Temperatures and Snow Cover at 5 Sites

A1.5.1 Churchill Permafrost Profile

This group of graphs show mean temperature profiles of ground temperatures recorded at Churchill. The 1978 and 1983 profiles were computed from 12 and 11 sets of readings, respectively. By 1989, a datalogger was in place and the readings were taken as follows:

interval	depth (metres)
hourly	0.5
2-hourly	1.2, 2.0, 2.7
6-hourly	3.5, 4.3, 5.8, 7.3
daily	8.8, 10.4, 11.9, 14.9

The smooth curve and the slightly warmer mean temperatures for 1989 can probably be attributed to the larger size of the sample. It is also important to note that the range of the mean temperature at 14.9 metres and at 0.5 metres is less than 1.5°C.

A1.5.2 Snow Depth at Thompson, Kuujjuaq, Baker Lake, and Pickle Lake

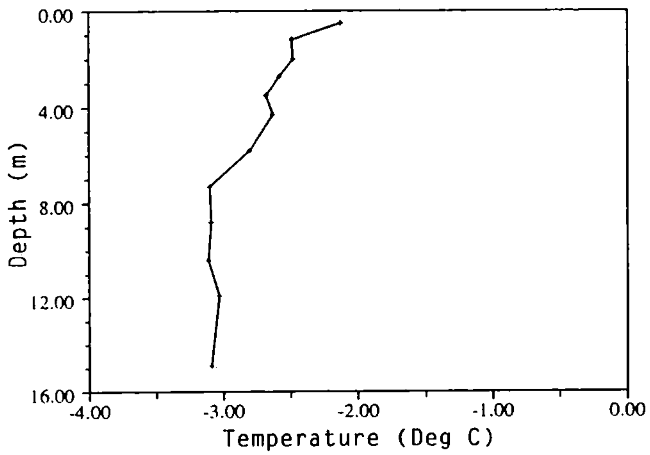
This set shows means, extremes, and standard deviations of snow depths observed at the four sites. The snow disappears from the ground by the beginning of June, except at Baker Lake, where it remains until early July. The ground usually remains snow free until early September at all sites. Light flurries can occur at any time of year.

Maximum snow cover occurs at Thompson in February. An extreme high of 155 cm was recorded in February 1977. Maximum cover occurs in March at Pickle Lake and Kuujjuaq, and April at Baker Lake.

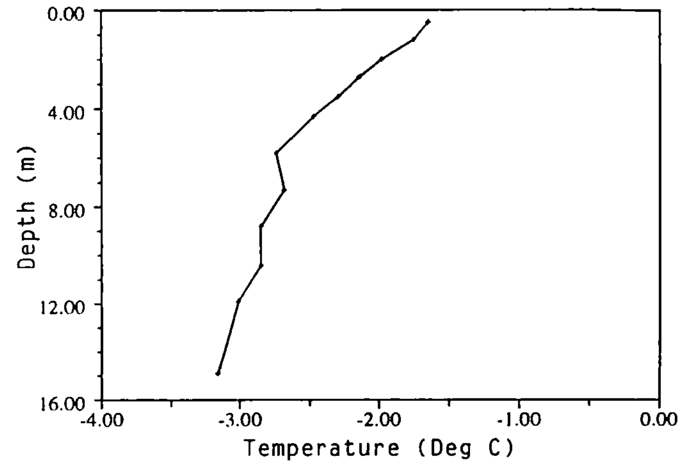
A1.5.3 to A1.5.6 Ground Temperatures at Thompson, Kuujjuaq, Baker Lake, and Pickle Lake

This set of graphs contains monthly means, extremes, and standard deviations of temperatures recorded at 0.1 metre to as deep as 3 metres. The temperature range decreases as the depth of the sensor in the soil increases. The standard deviations also decrease with depth.

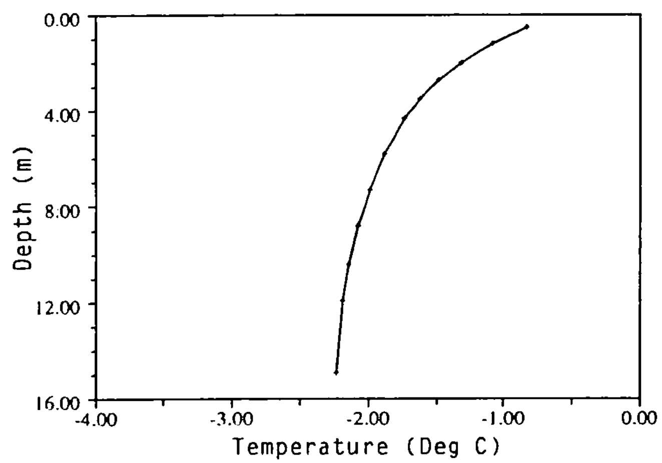
CHURCHILL PERMAFROST PROFILE
(Averaged: Day 23 - Day 348, 1978)



CHURCHILL PERMAFROST PROFILE
(Averaged: Day 50 - Day 350, 1983)

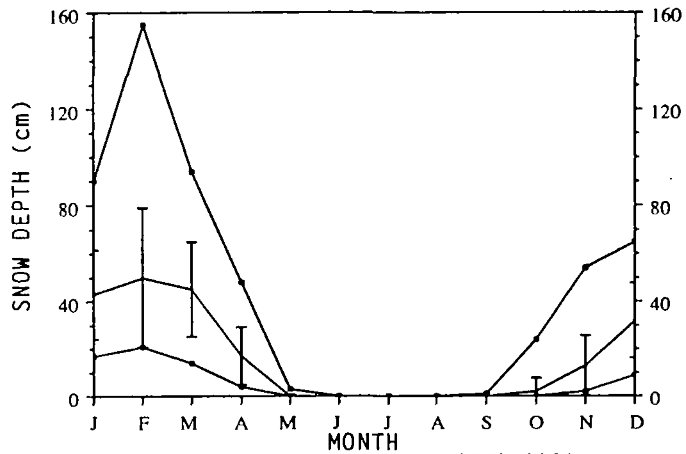


CHURCHILL PERMAFROST PROFILE
(Averaged: Day 26, 1988 - Day 64, 1989)



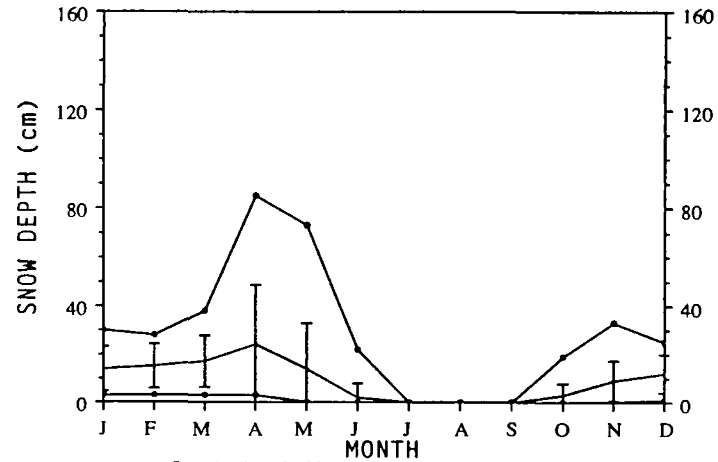
APPENDIX 1.5.1 Churchill Permafrost Profile

THOMPSON A, MANITOBA
AM Snow Depth: Max, Min, Mean & Standard Deviation



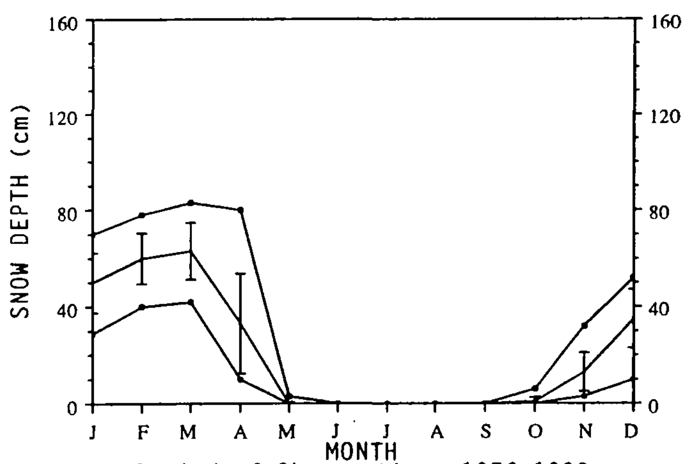
Period of Observation: 1970-1991

BAKER LAKE, N.W.T
AM Snow Depth: Max, Min, Mean & Standard Deviation



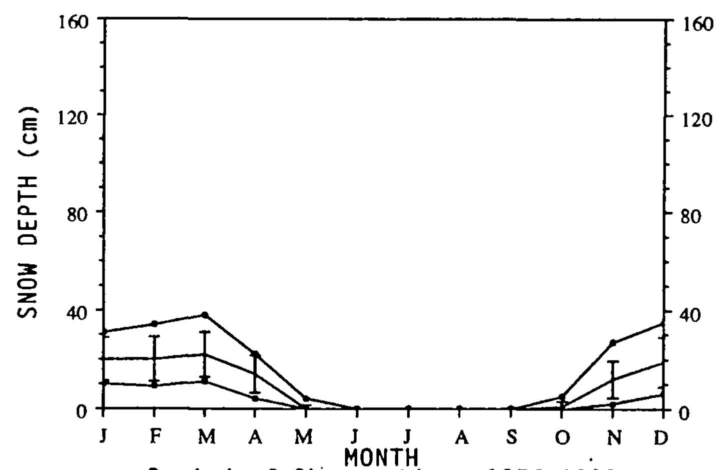
Period of Observation: 1964-1980

PICKLE LAKE, ONTARIO
AM Snow Depth: Max, Min, Mean & Standard Deviation



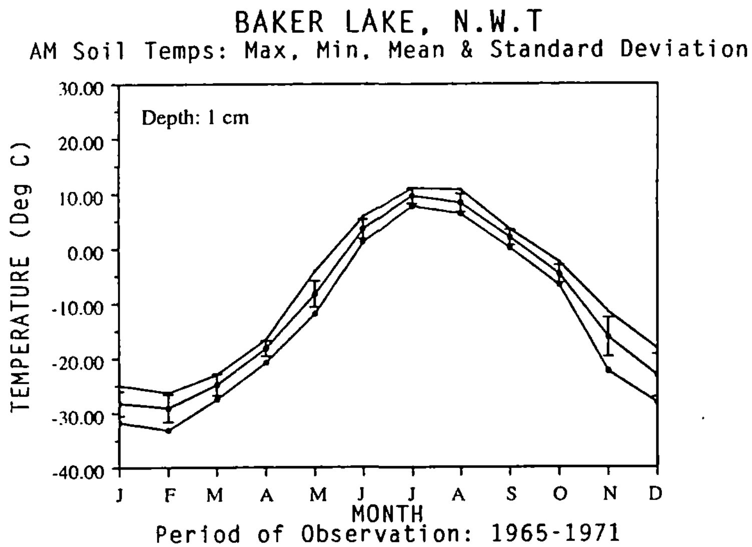
Period of Observation: 1976-1990

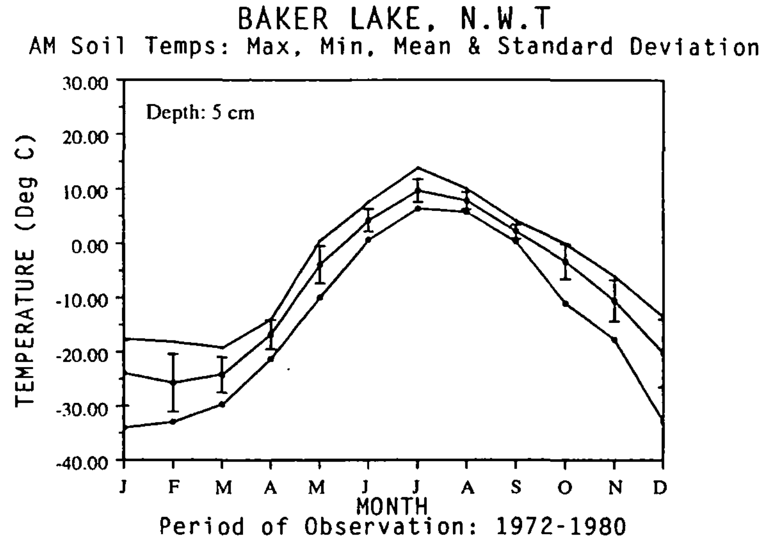
KUUJJUAQ UA, QUEBEC
AM Snow Depth: Max, Min, Mean & Standard Deviation

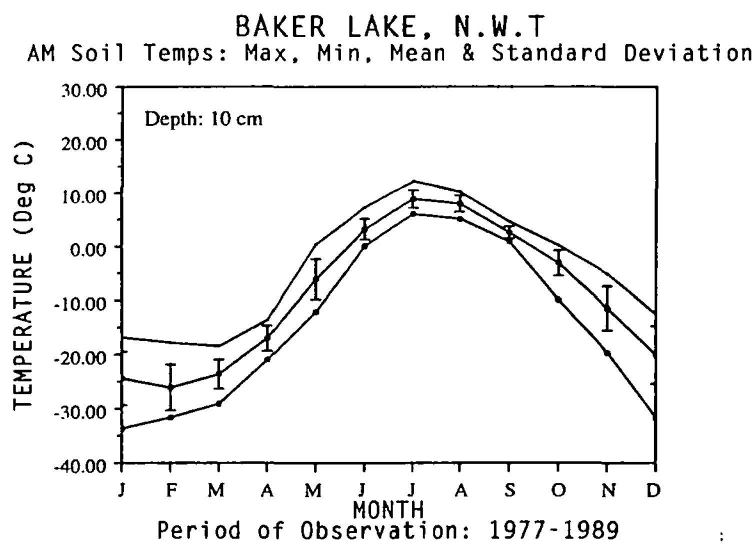


Period of Observation: 1976-1990

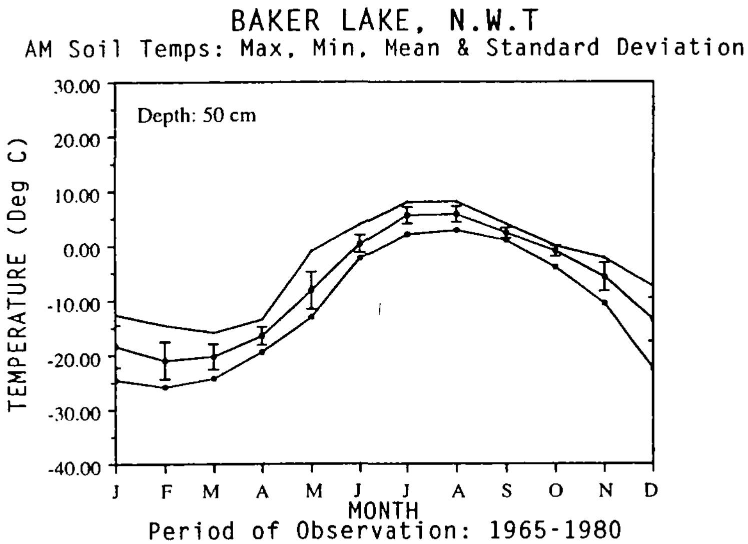
APPENDIX 1.5.2 Snow Depth

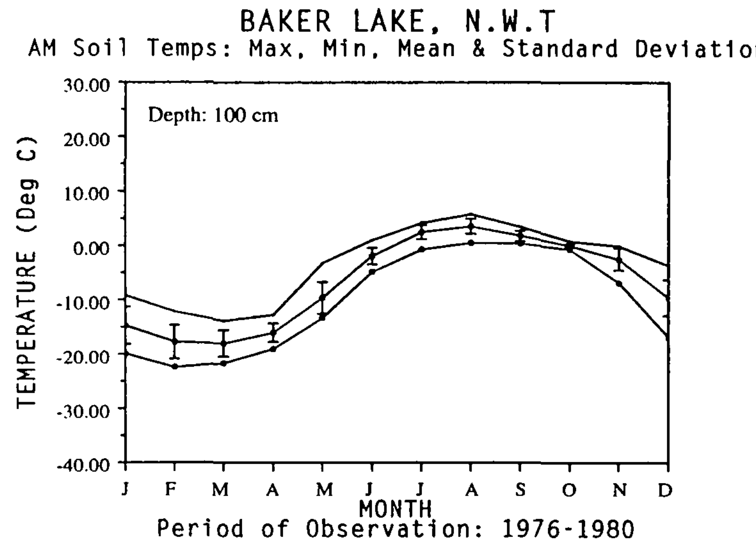


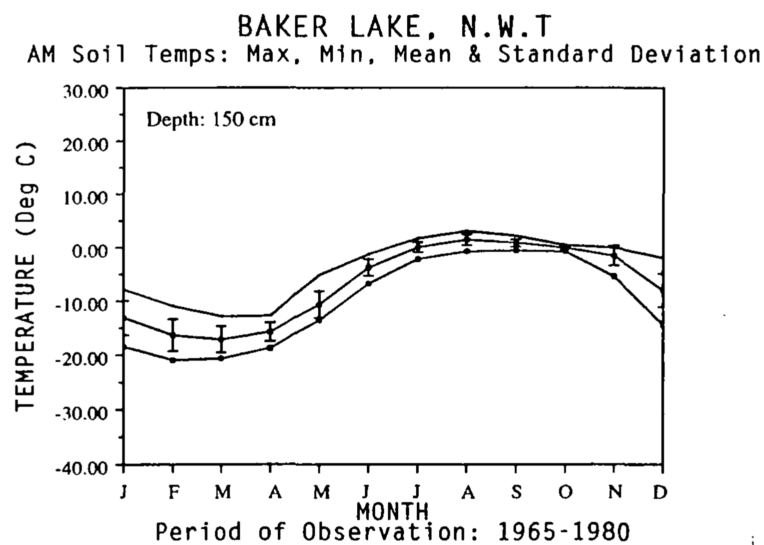




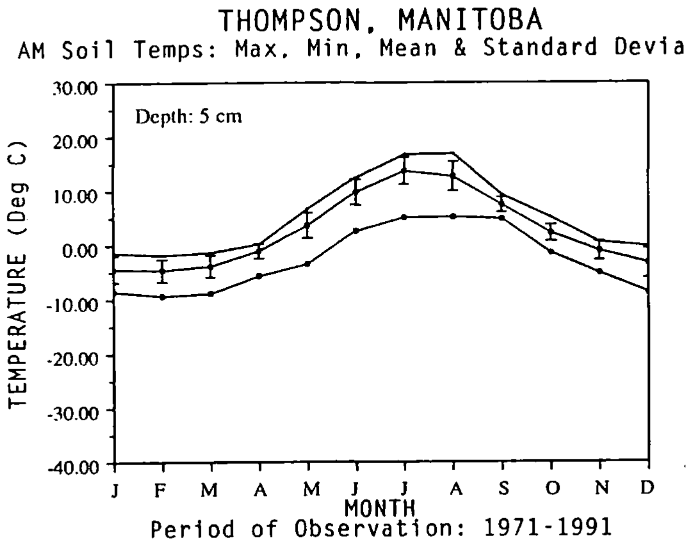
APPENDIX 1.5.3A Baker Lake Soil Temperatures

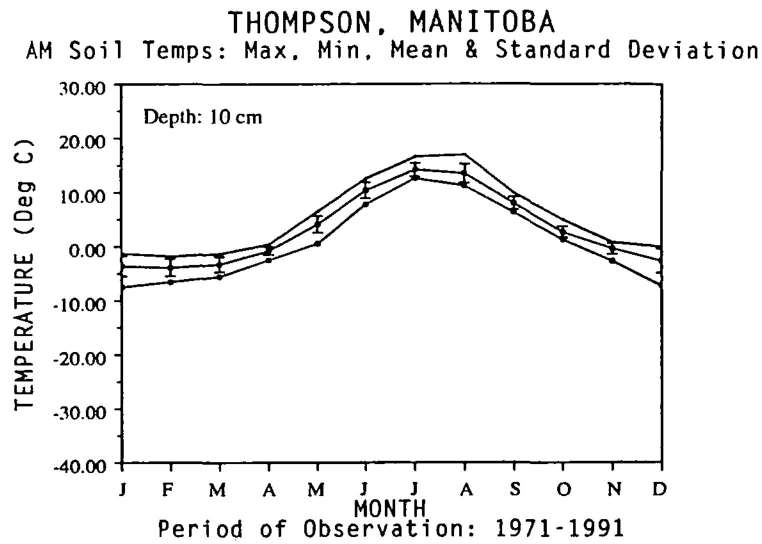


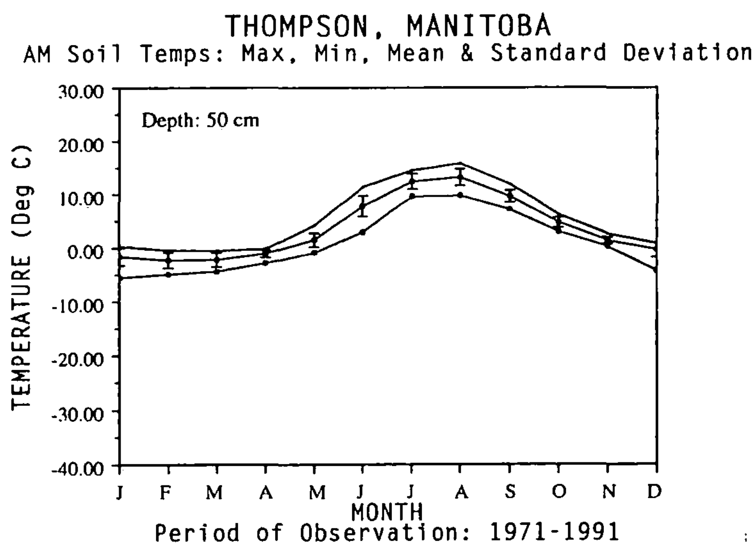




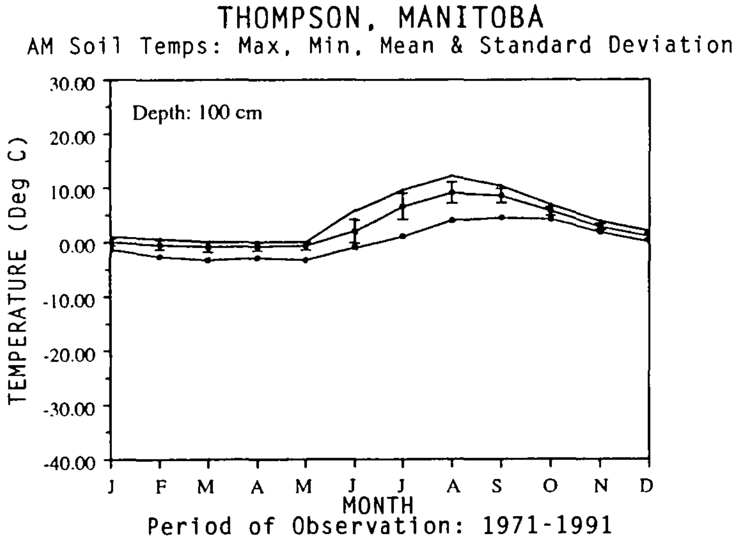
APPENDIX 1.5.3B Baker Lake Soil Temperatures

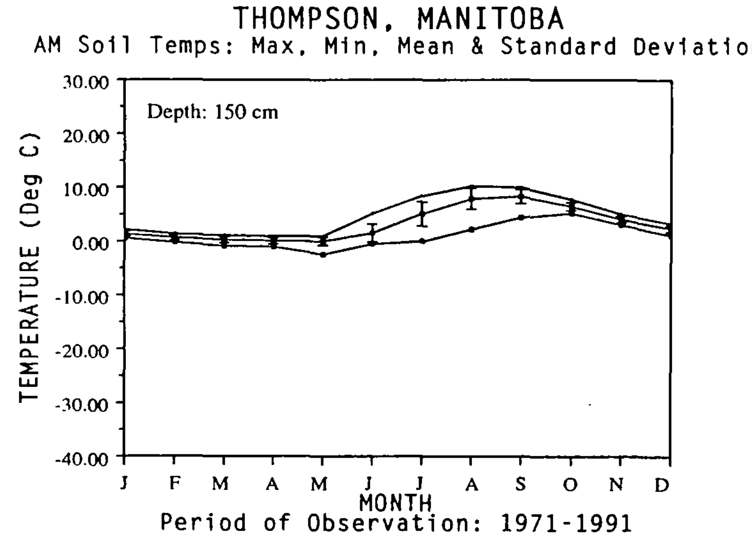


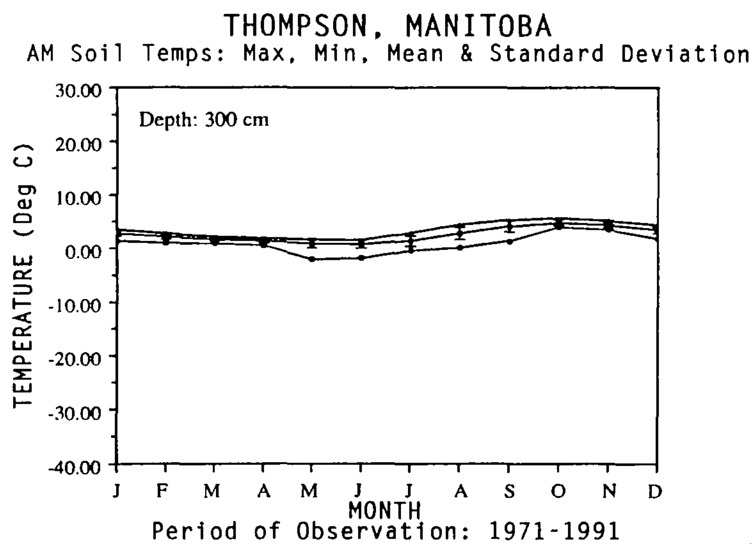




APPENDIX 1.5.4A Thompson Soil Temperatures

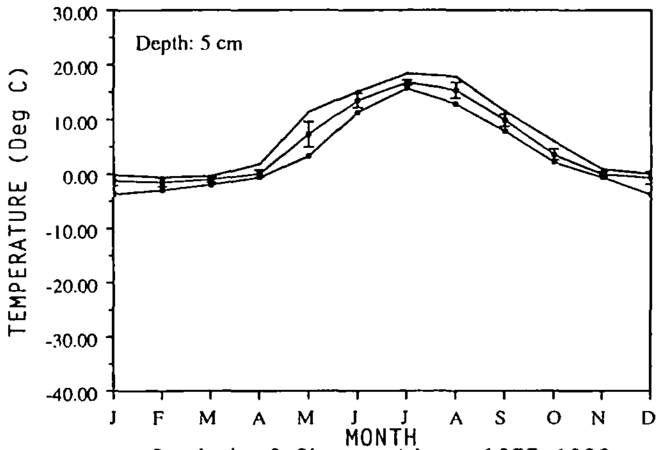






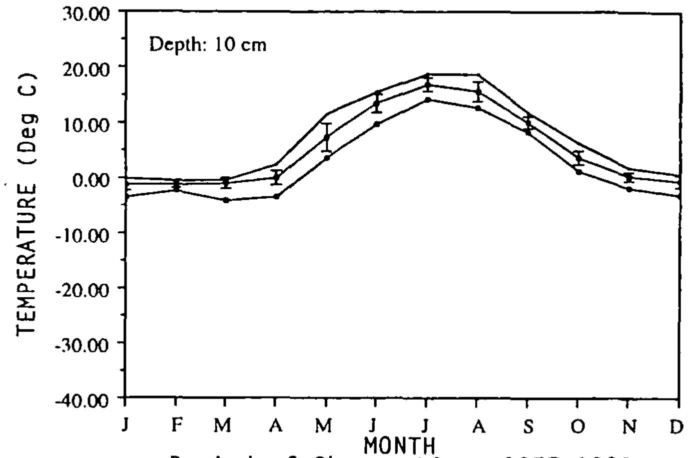
APPENDIX 1.5.4B Thompson Soil Temperatures

PICKLE LAKE, ONTARIO
AM Soil Temps: Max, Min, Mean & Standard Deviation



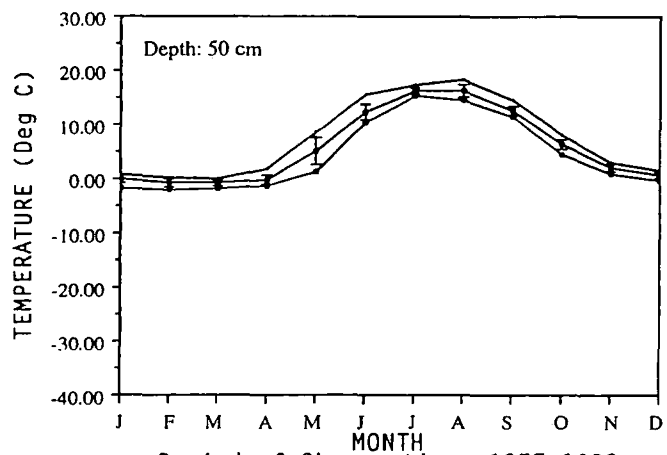
Period of Observation: 1977-1989

PICKLE LAKE, ONTARIO
AM Soil Temps: Max, Min, Mean & Standard Deviation



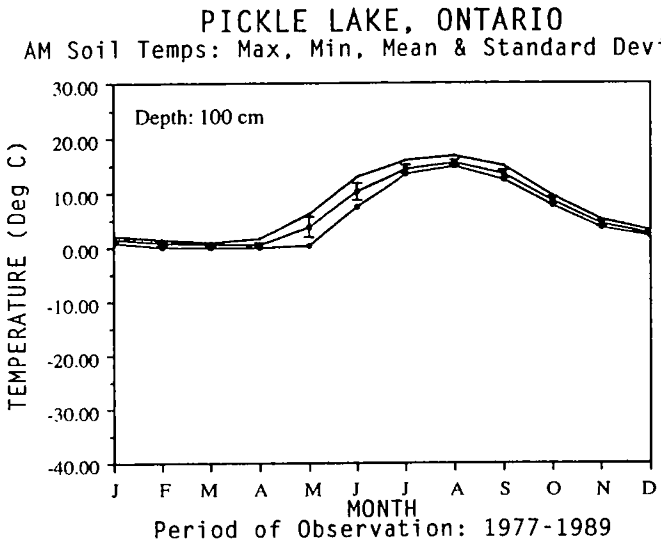
Period of Observation: 1977-1989

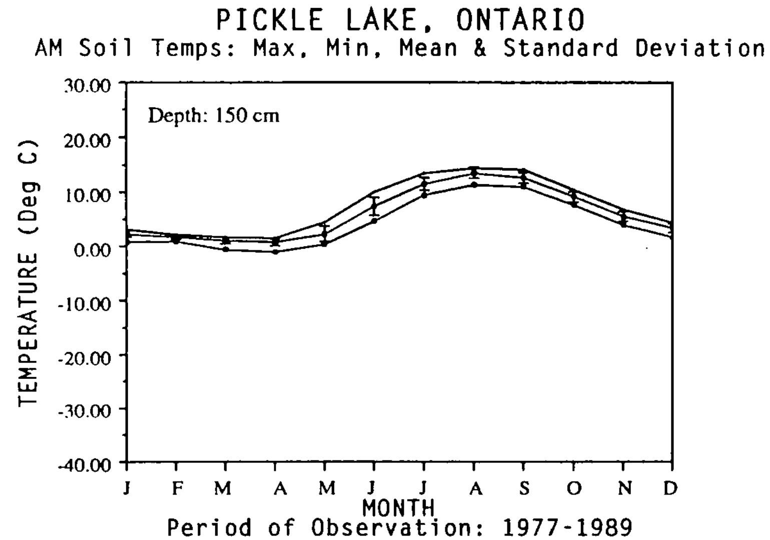
PICKLE LAKE, ONTARIO
AM Soil Temps: Max, Min, Mean & Standard Deviation

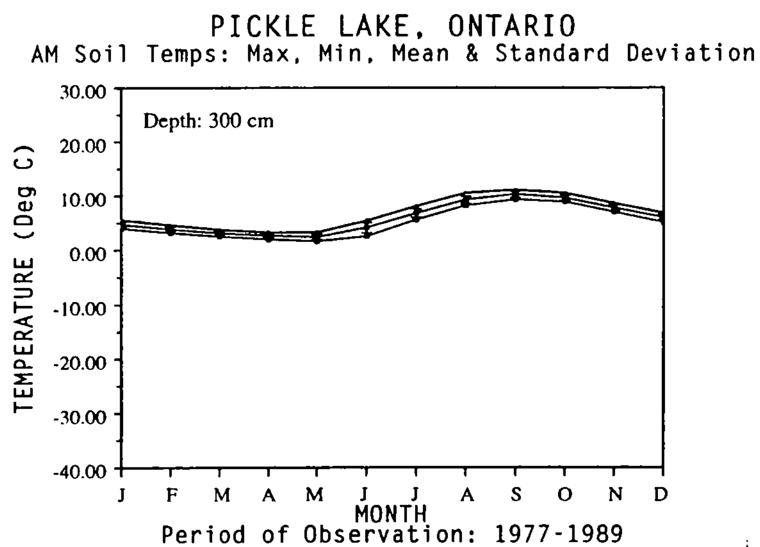


Period of Observation: 1977-1989

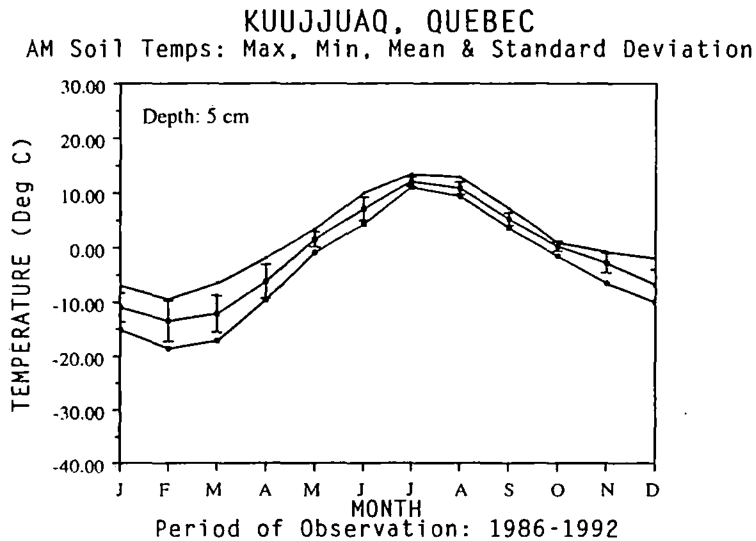
APPENDIX 1.5.5A Pickle Lake Soil Temperatures

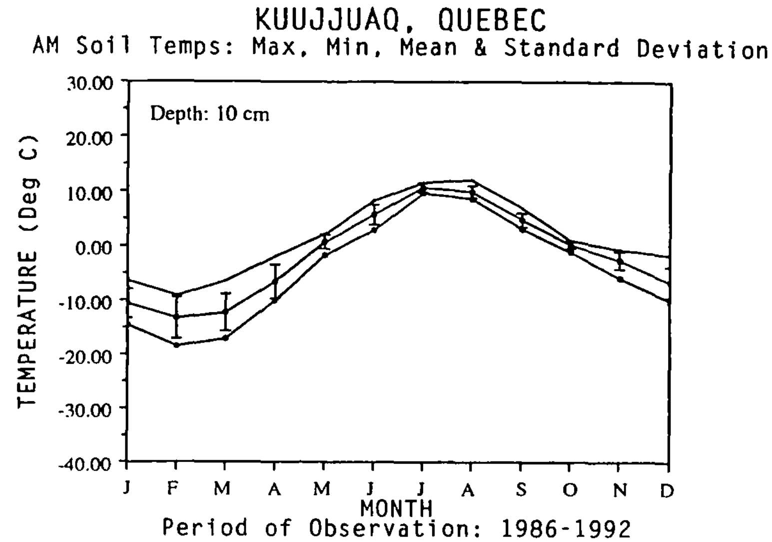


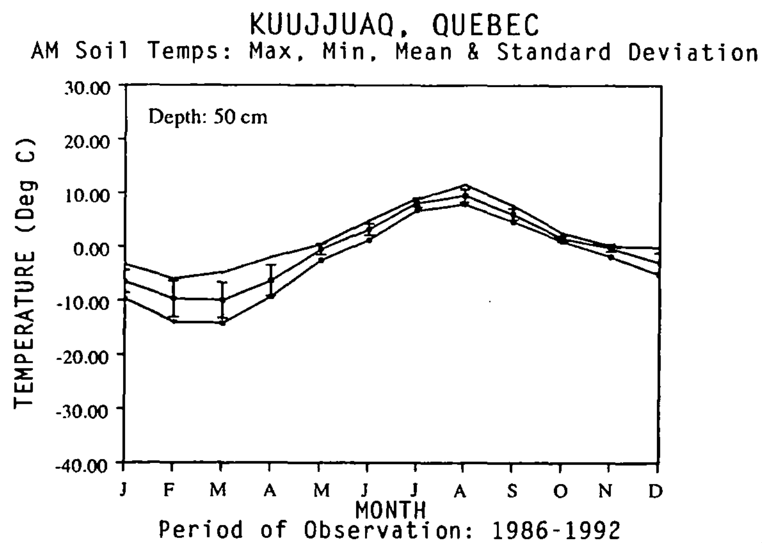




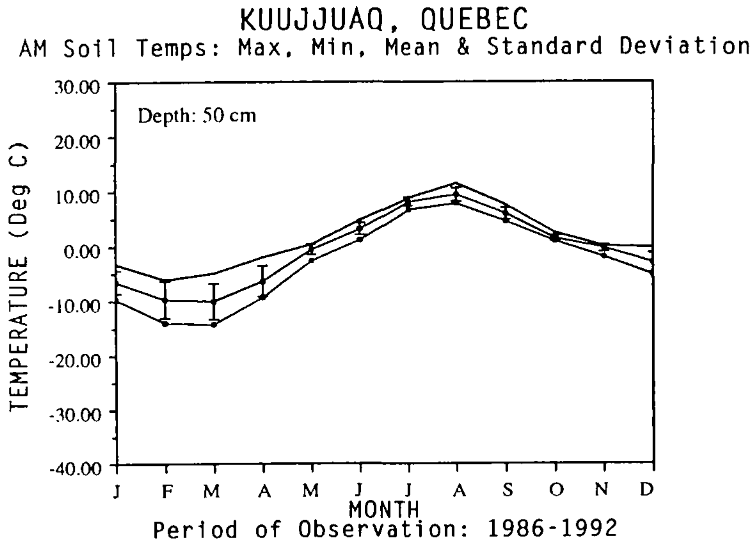
APPENDIX 1.5.5B Pickle Lake Soil Temperatures

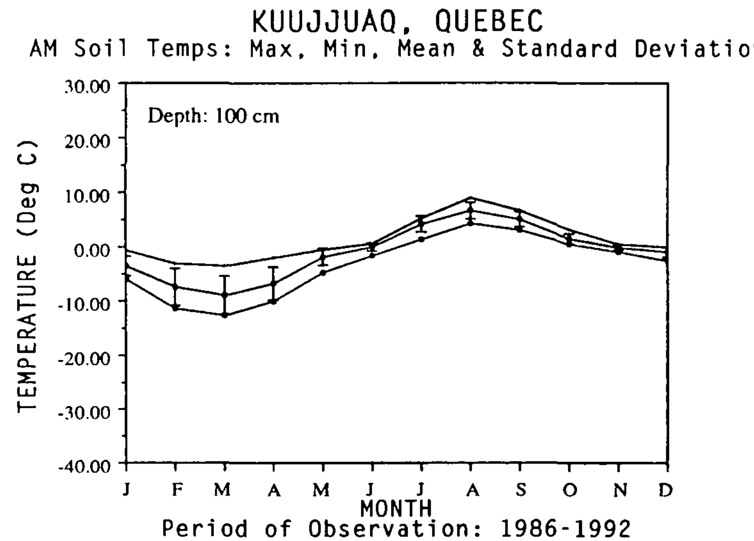


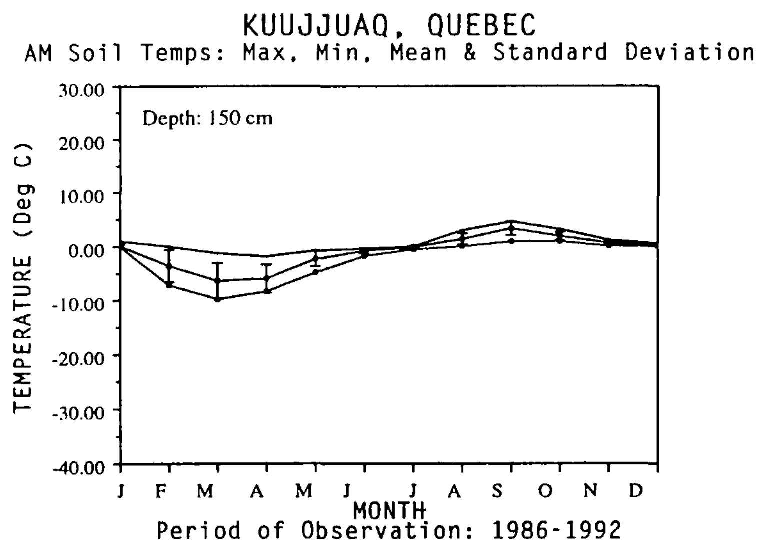




APPENDIX 1.5.6A Kuujjuaq Soil Temperatures





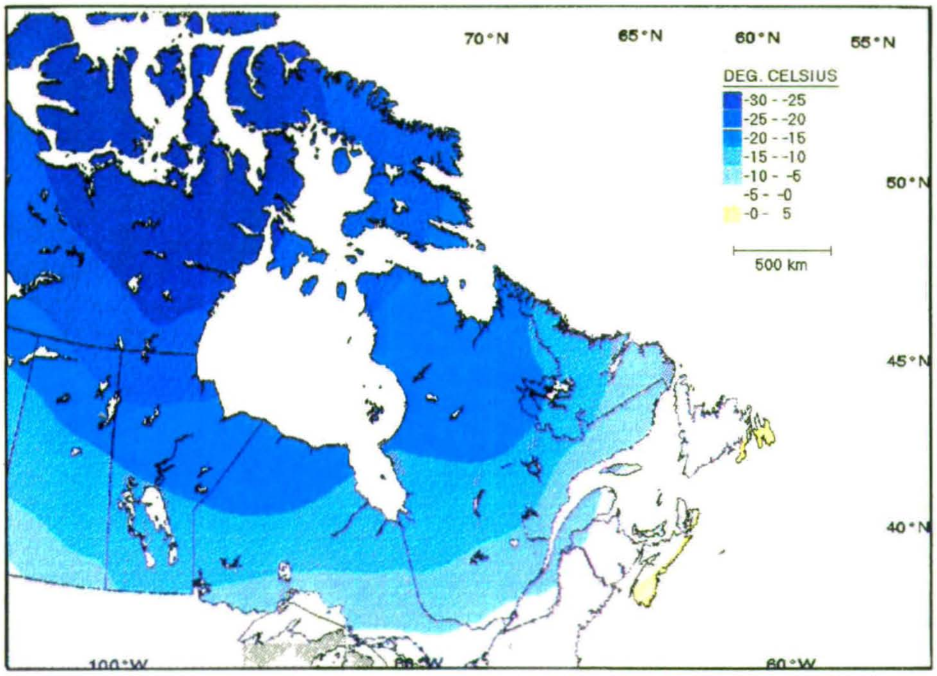


APPENDIX 1.5.6B Kuujjuaq Soil Temperatures

APPENDIX 2. CLIMATIC VARIABILITY AND TRENDS

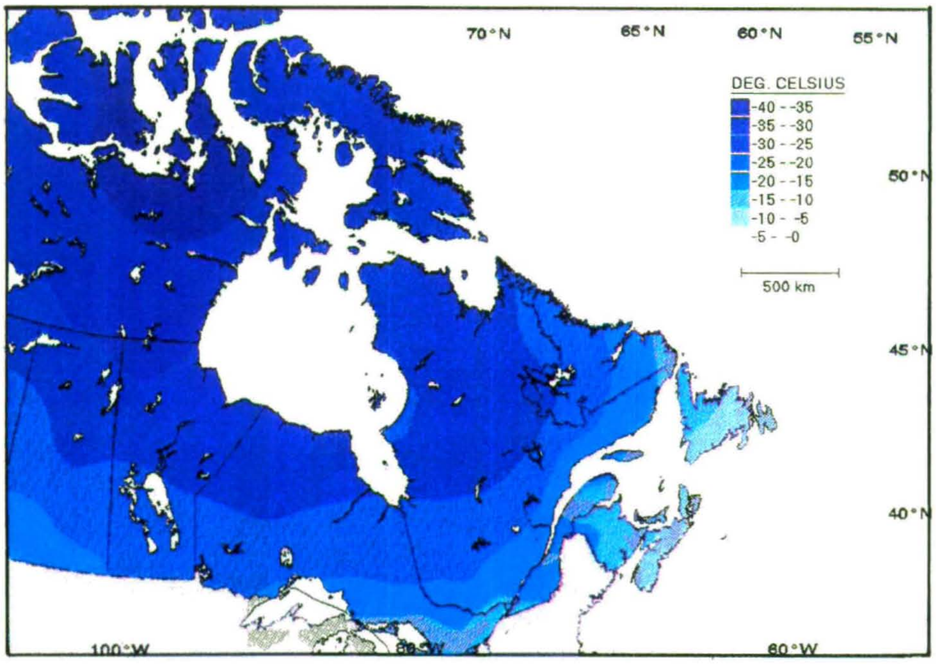
A2.2 Average Winter and Summer Temperature and Precipitation.

Appendix 2.2.1 Winter average maximum temperature, 1951-80.



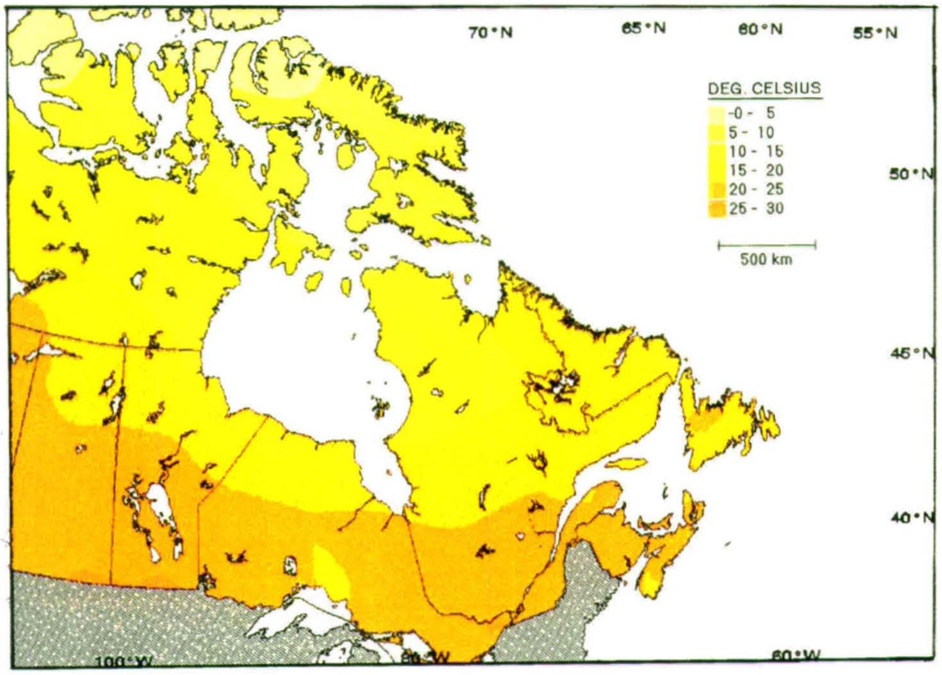
Source: Environment Canada

Appendix 2.2.2 Winter average minimum temperature, 1951-80.



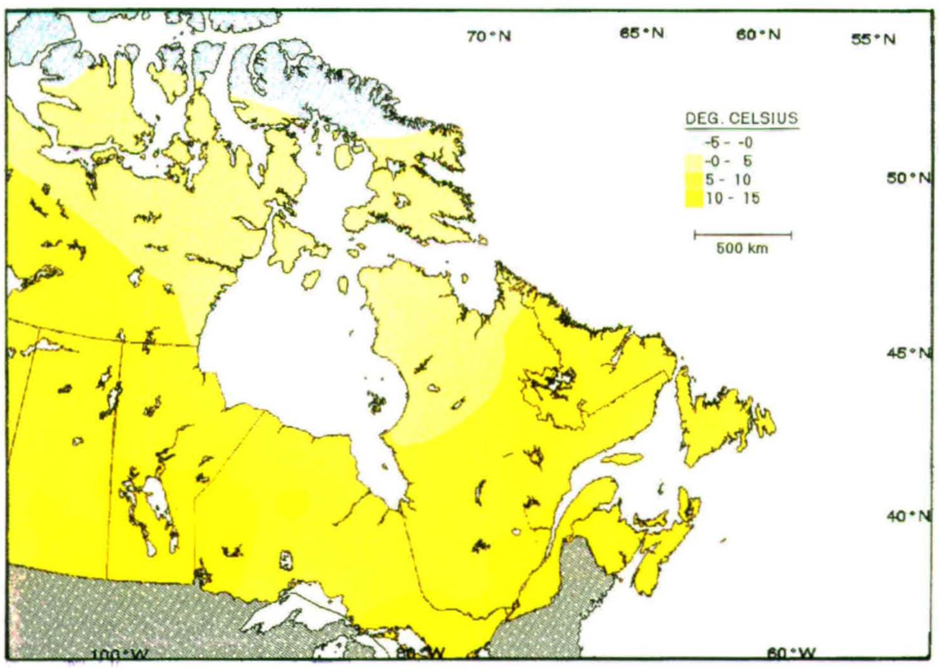
Source: Environment Canada

Appendix 2.2.3 Summer average maximum temperature, 1951-80.



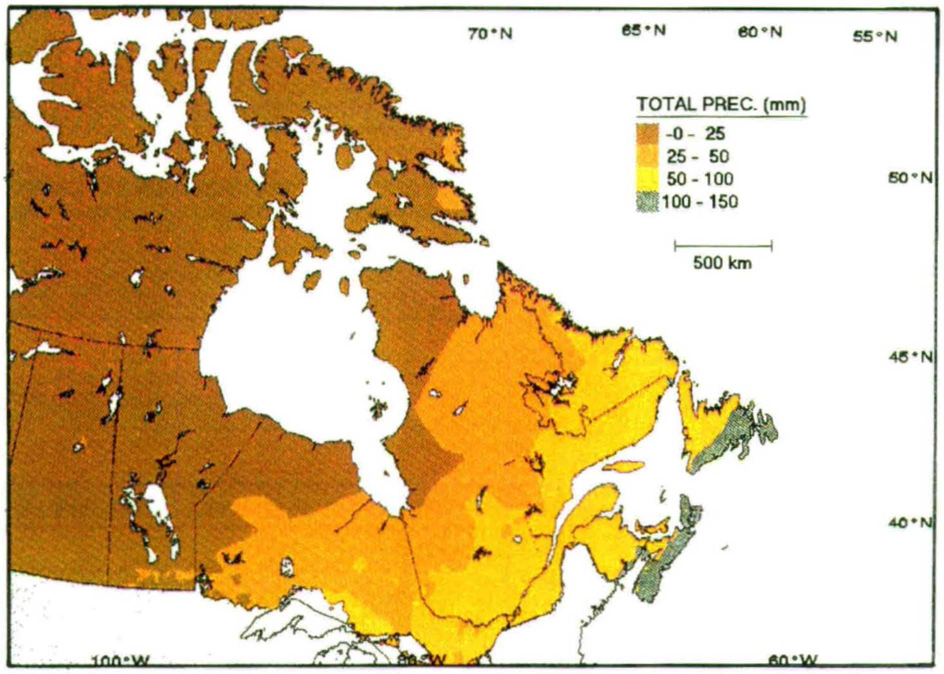
Source: Environment Canada

Appendix 2.2.4 Summer average minimum temperature, 1951-80.



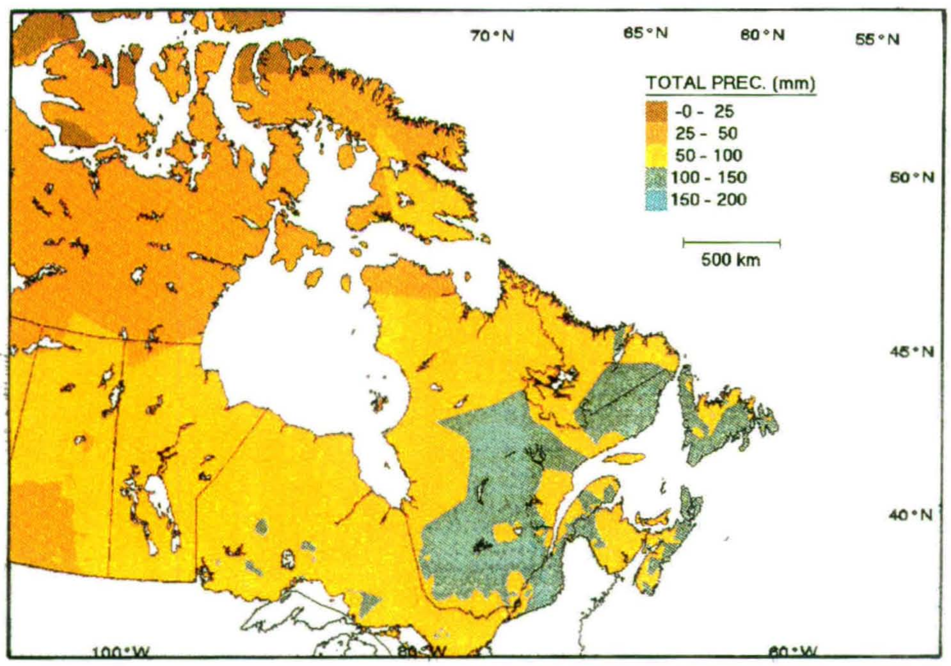
Source: Environment Canada

Appendix 2.2.5 February average total precipitation, 1951-80.



Source: Environment Canada

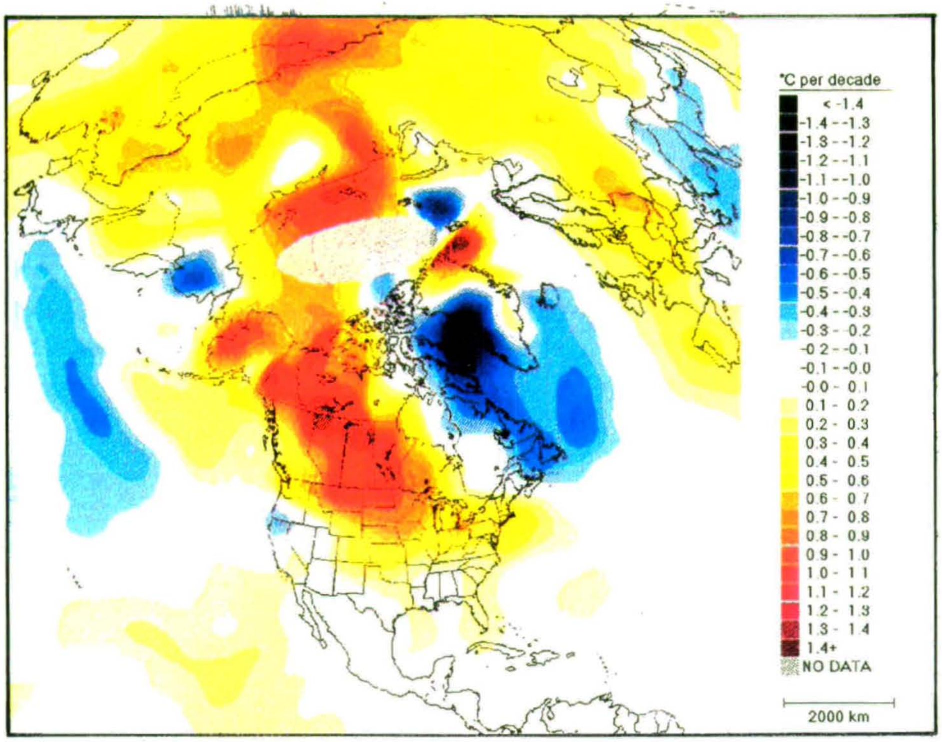
Appendix 2.2.6 August average total precipitation, 1951-80.



Source: Environment Canada

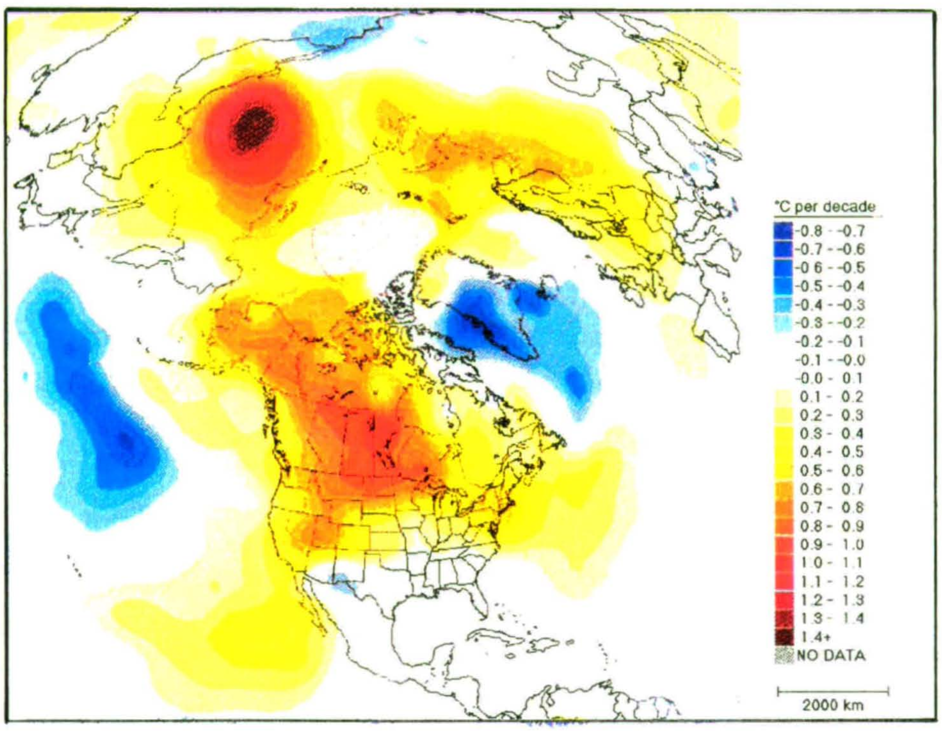
A2.5 Seasonal Trends in Temperature, Precipitation and Lake Ice.

Appendix 2.5.1 Winter surface temperature trends, 1961-90.



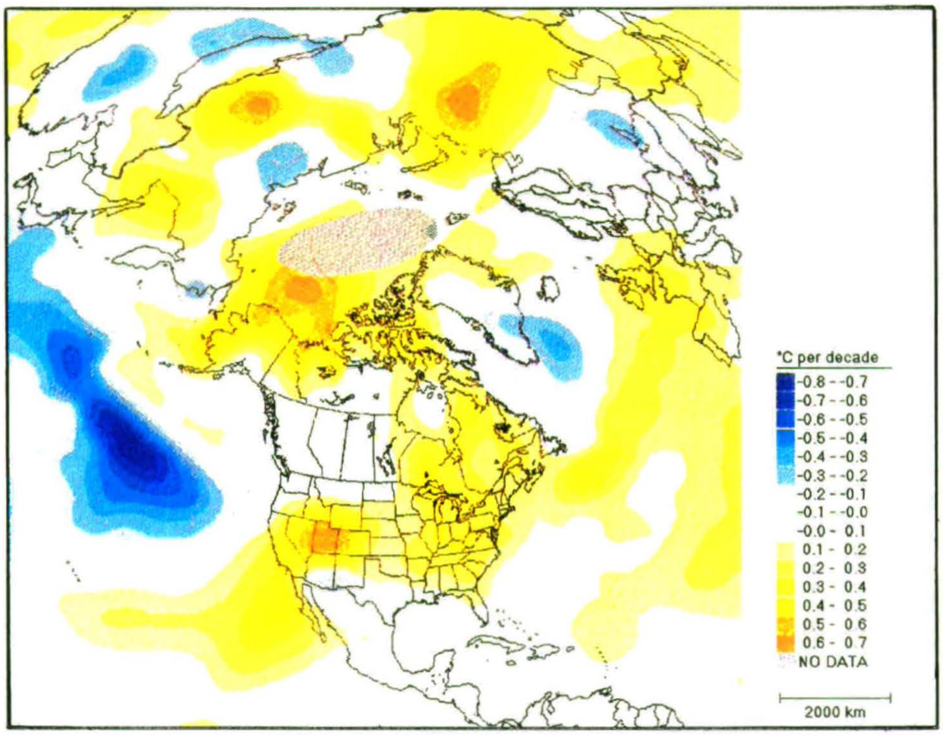
Source: Environment Canada

Appendix 2.5.2 Spring surface temperature trends, 1961-90.



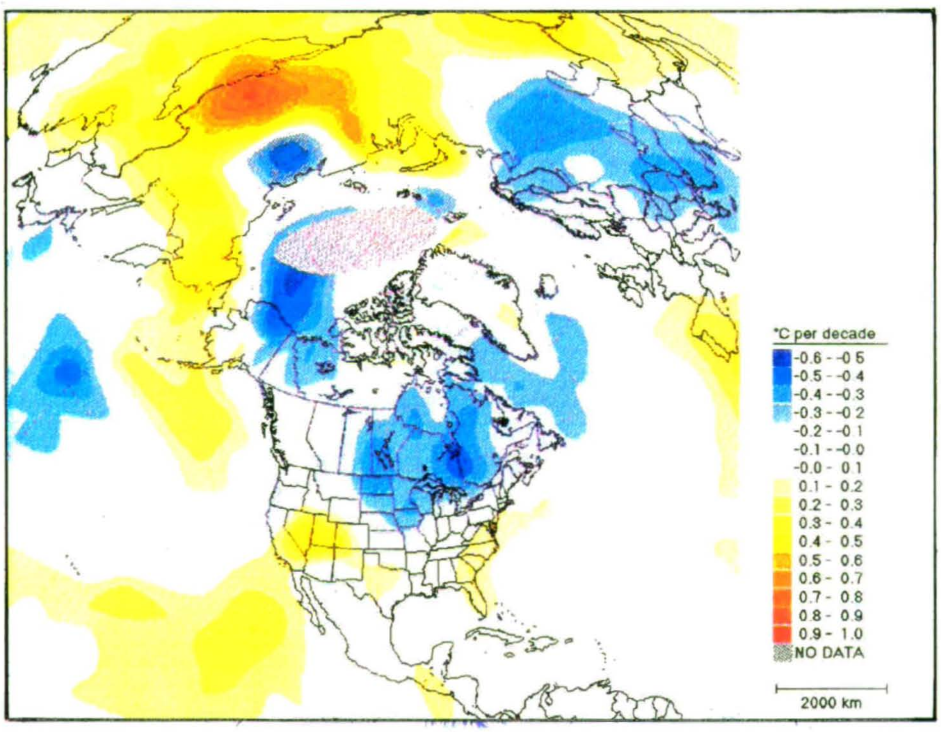
Source: Environment Canada

Appendix 2.5.3 Summer surface temperature trends, 1961-90.



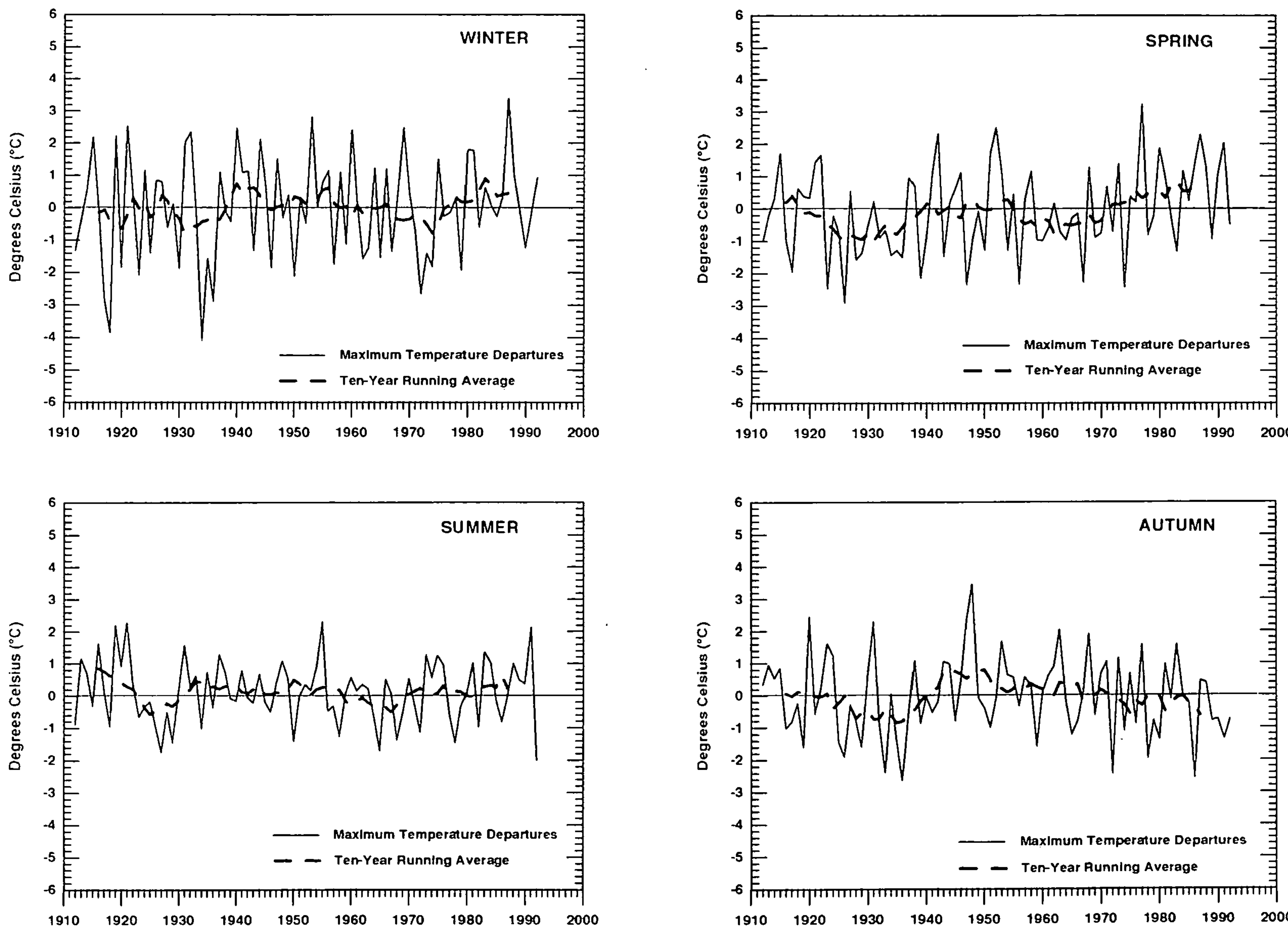
Source: Environment Canada

Appendix 2.5.4 Autumn surface temperature trends, 1961-90.



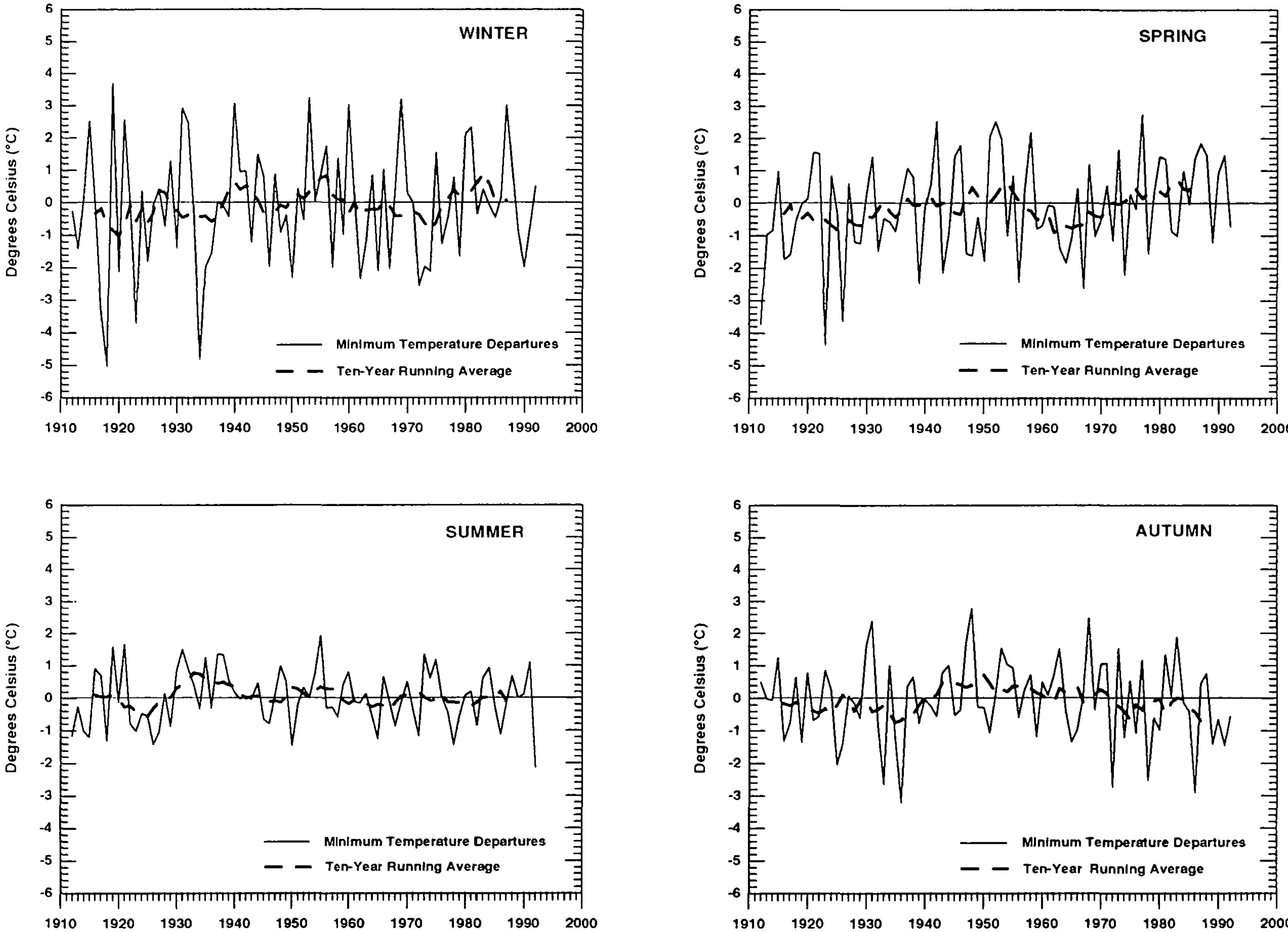
Source: Environment Canada

Appendix 2.5.5. Seasonal mean daily maximum temperature departures from the 1951-80 average for the Hudson Bay Bioregion.



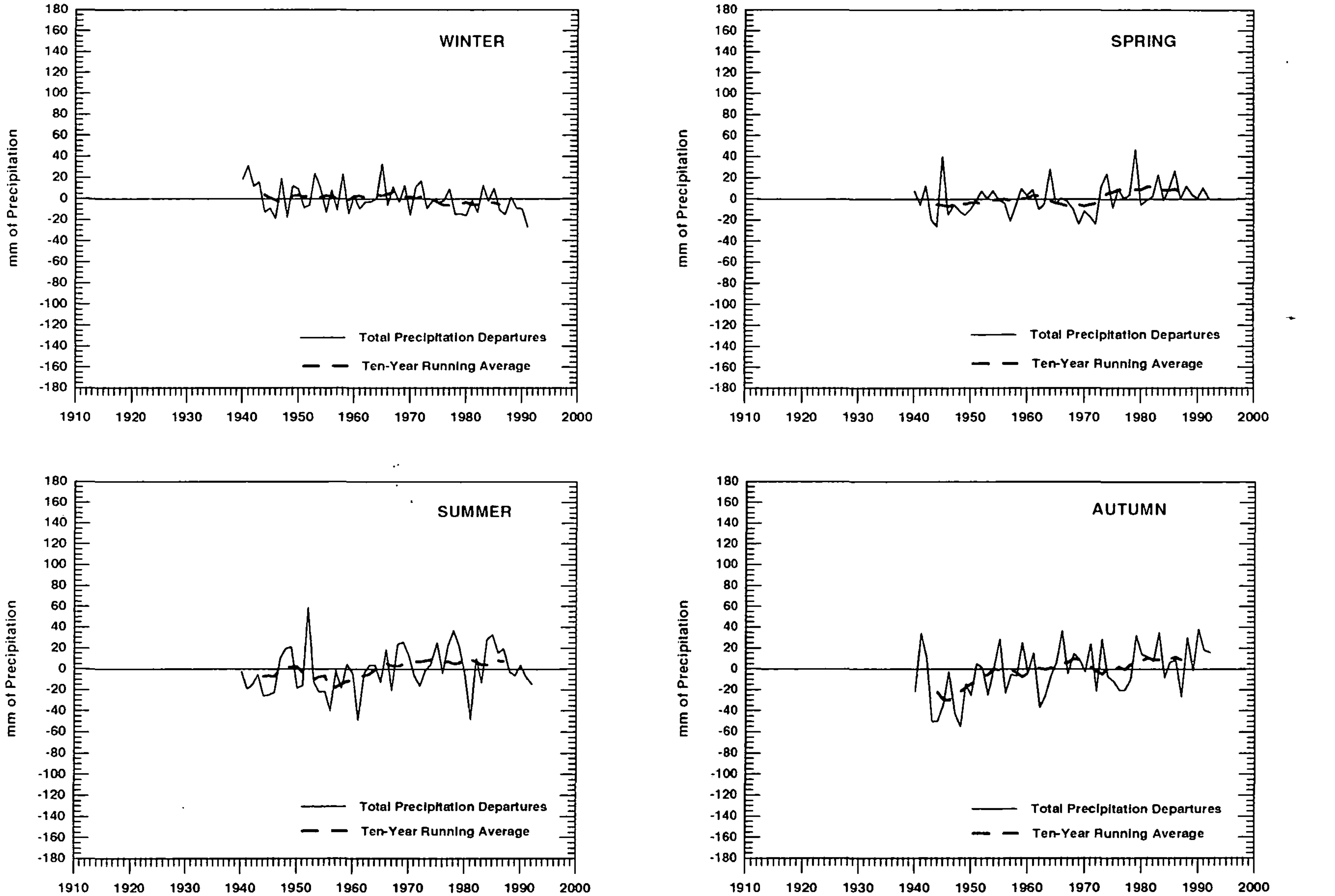
Source: Environment Canada

Appendix 2.5.6. Seasonal mean daily minimum temperature departures from the 1951-80 average for the Hudson Bay Bioregion.



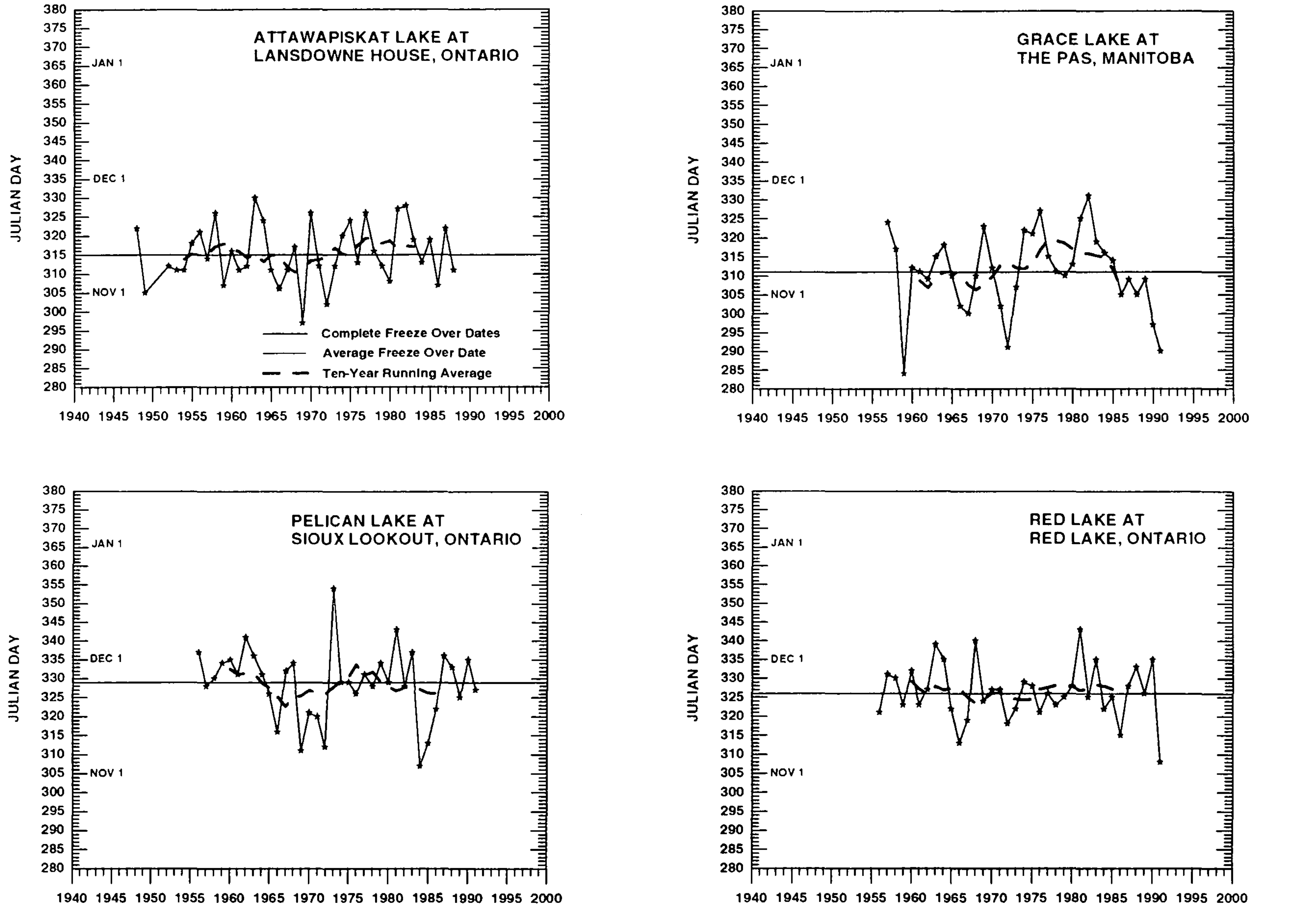
Source: Environment Canada

Appendix 2.5.7. Seasonal total precipitation departures from the 1951-80 average for the Hudson Bay Bioregion.



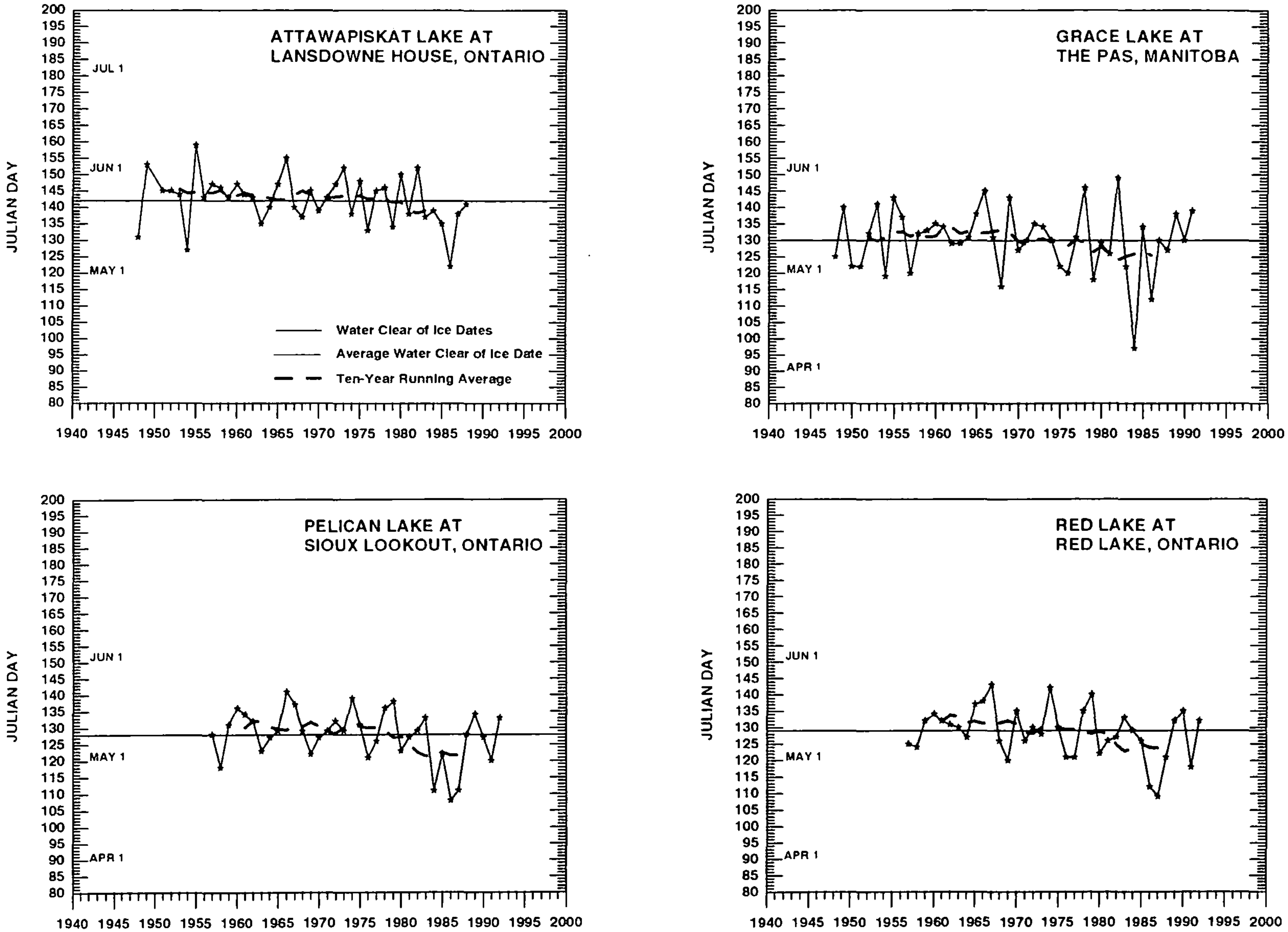
Source: Environment Canada

Appendix 2.5.8. Complete freeze over dates at lakes in the Hudson Bay Bioregion.



Source: Environment Canada

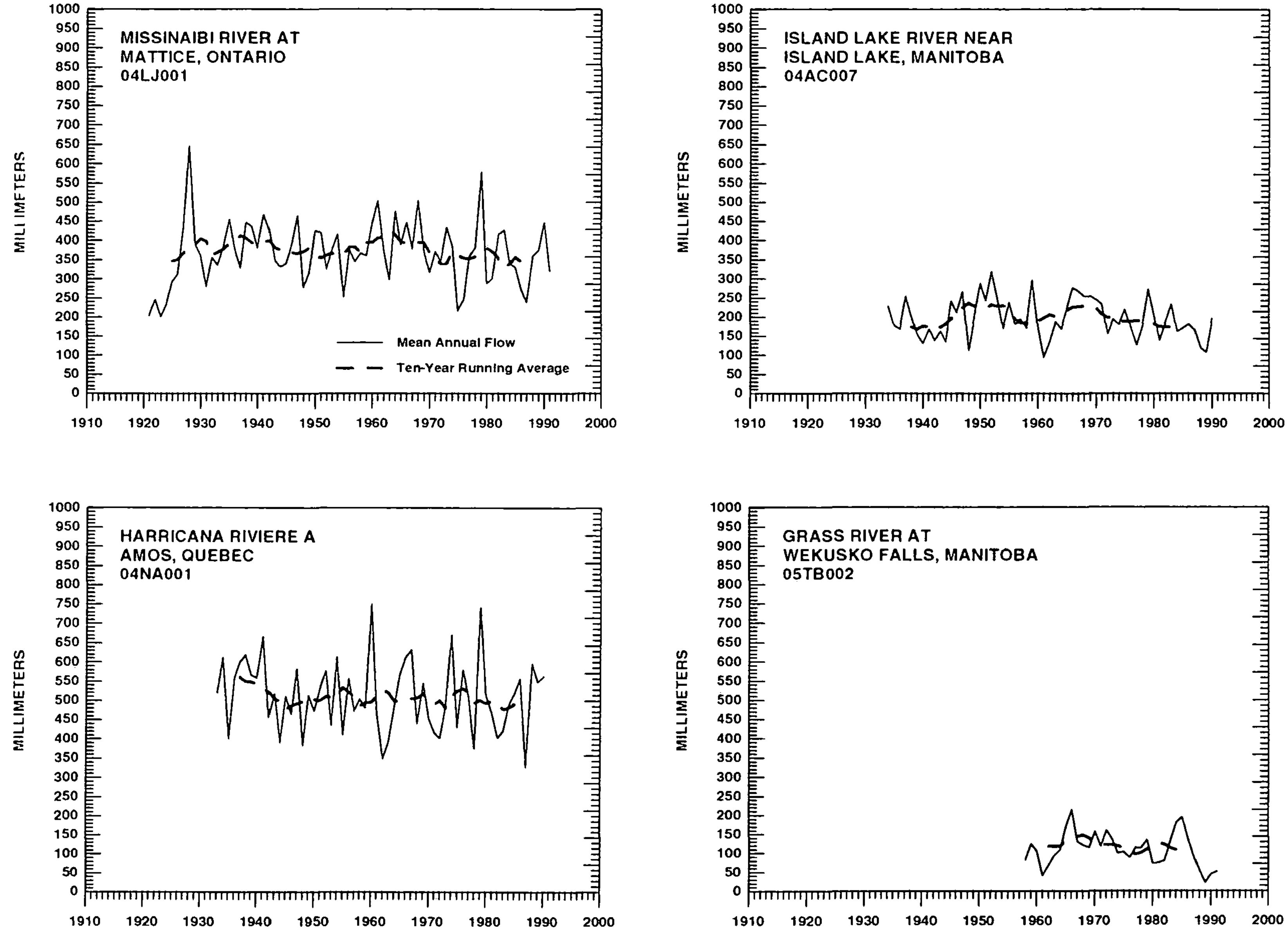
Appendix 2.5.9. Water clear of ice dates at lakes in the Hudson Bay Bioregion.



Source: Environment Canada

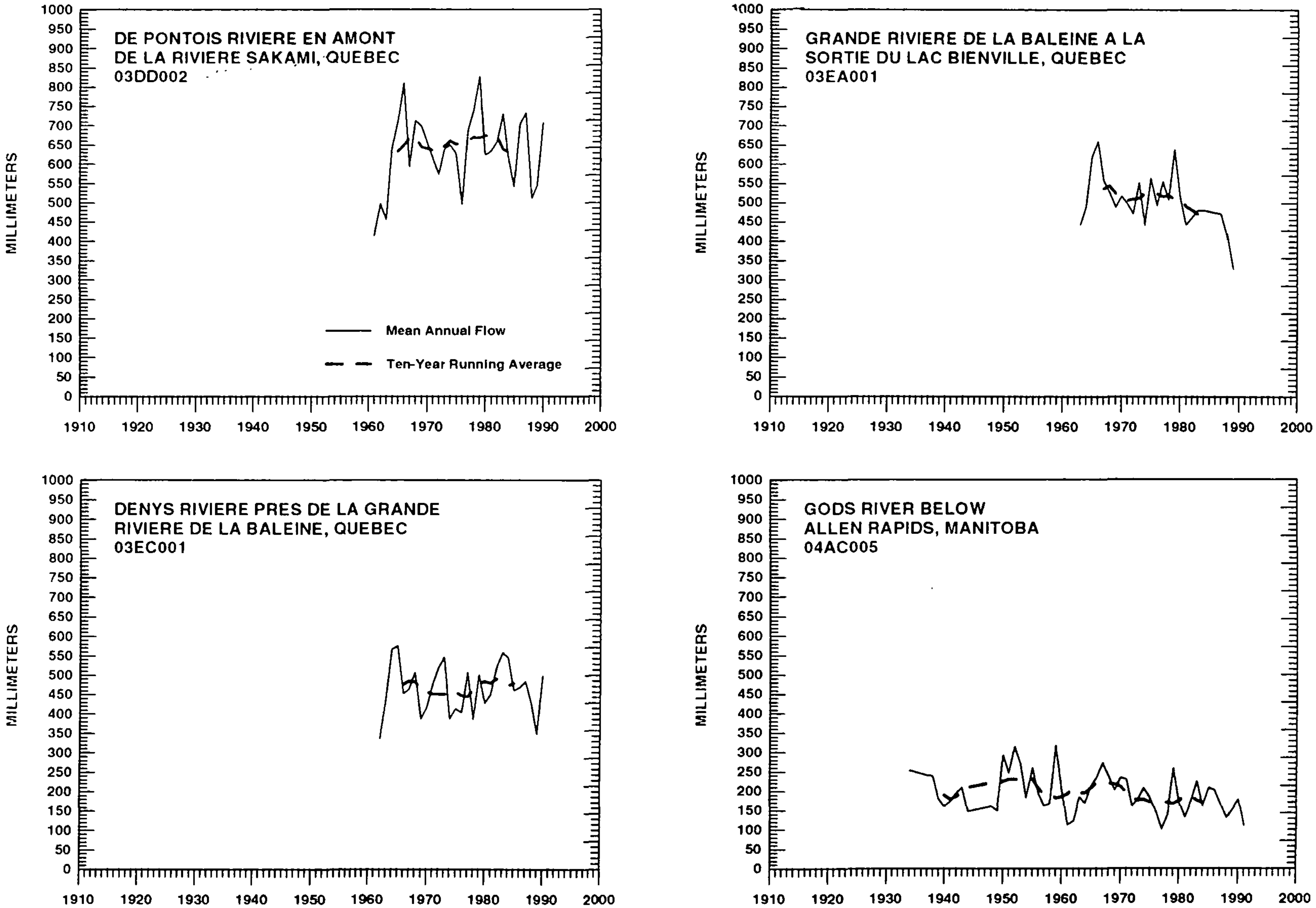
A2.6 Mean Annual Streamflow at 12 Sites.

Appendix 2.6(a). Mean annual streamflow at stations in the Hudson Bay Bioregion.



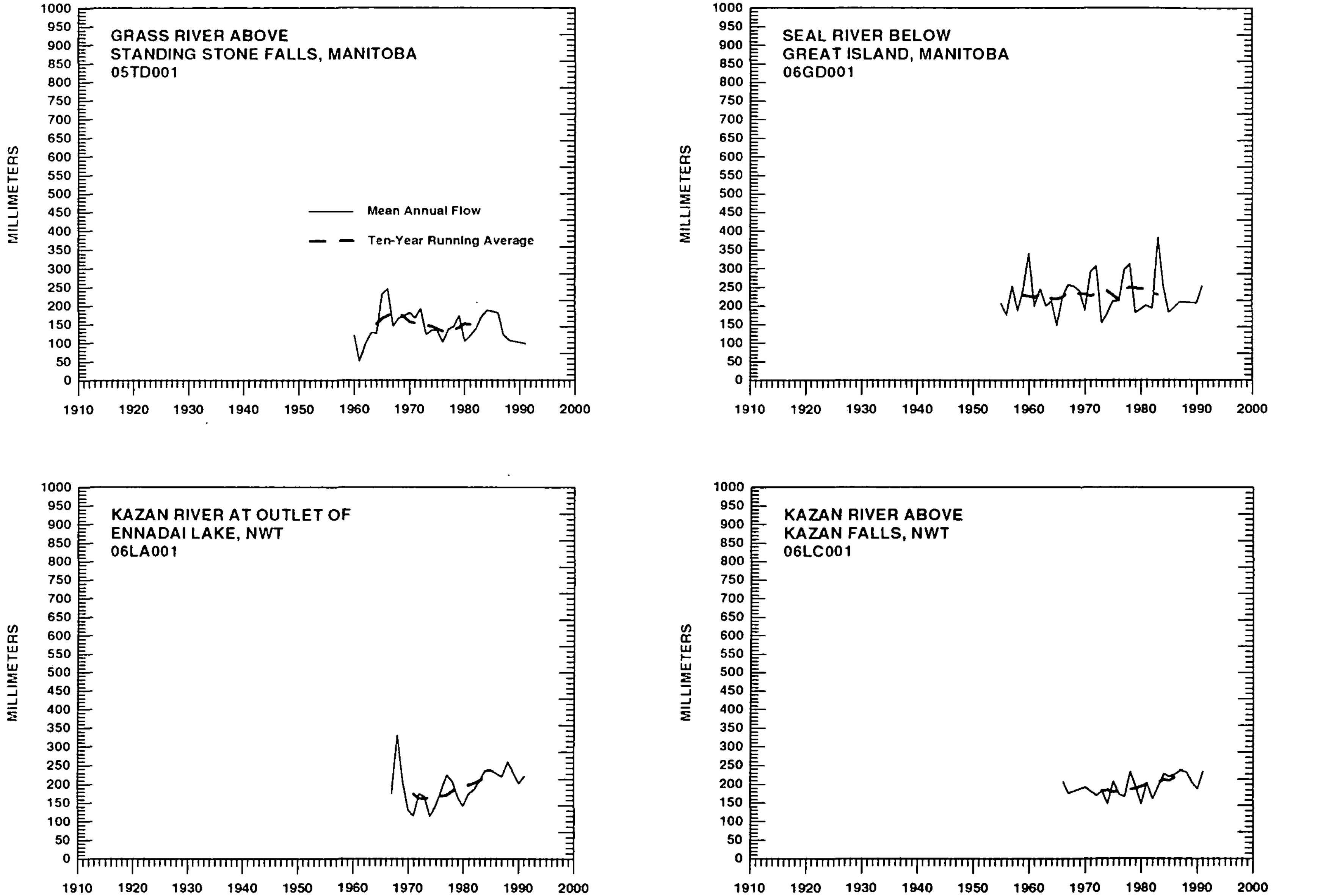
Source: Environment Canada

Appendix 2.6(b). Mean annual streamflow at stations in the Hudson Bay Bioregion.



Source: Environment Canada

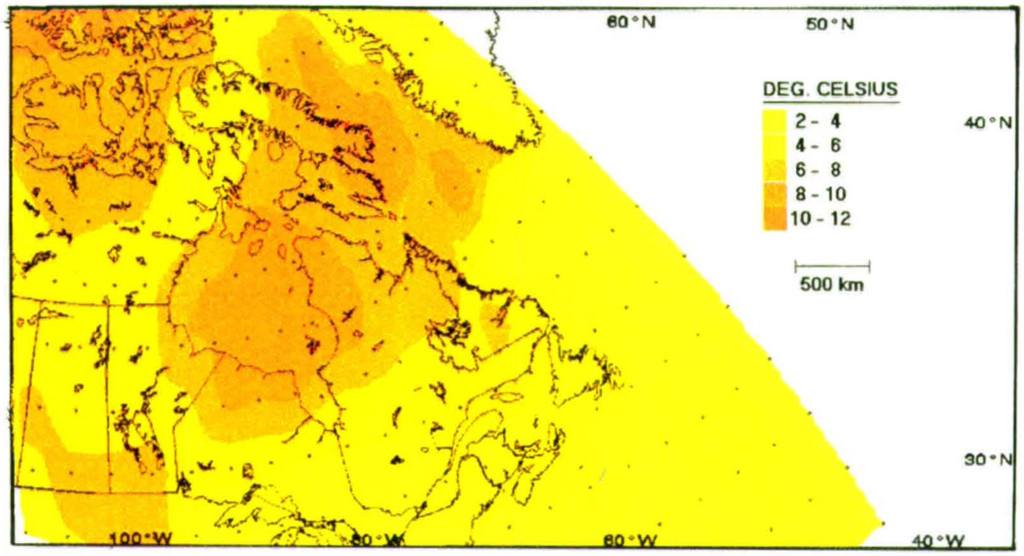
Appendix 2.6(c). Mean annual streamflow at stations in the Hudson Bay Bioregion.



Source: Environment Canada

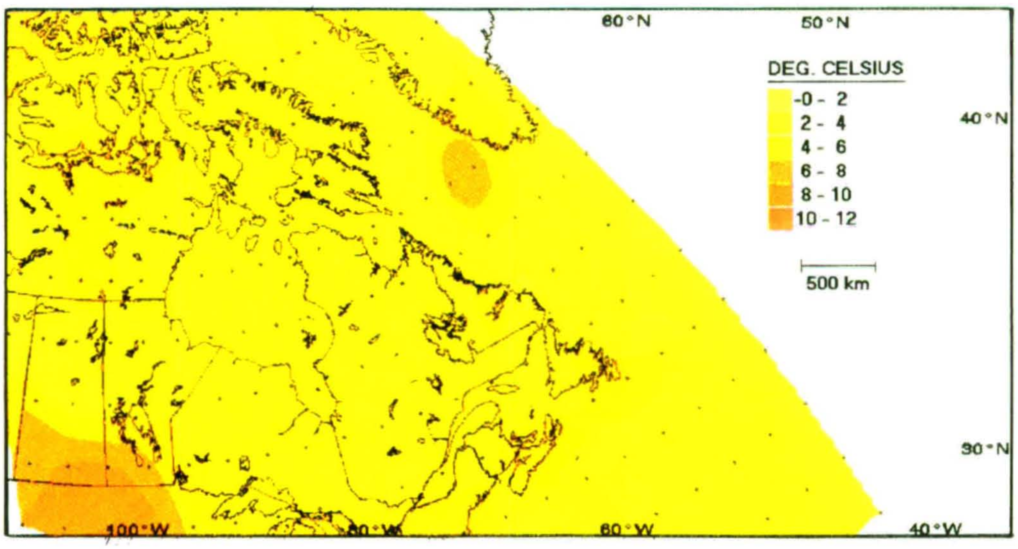
APPENDIX 3. SIMULATIONS OF PROJECTED CLIMATIC CHANGE OBTAINED FROM CCC GCM2.

Appendix 3.4.1 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for winter daily maximum temperature.



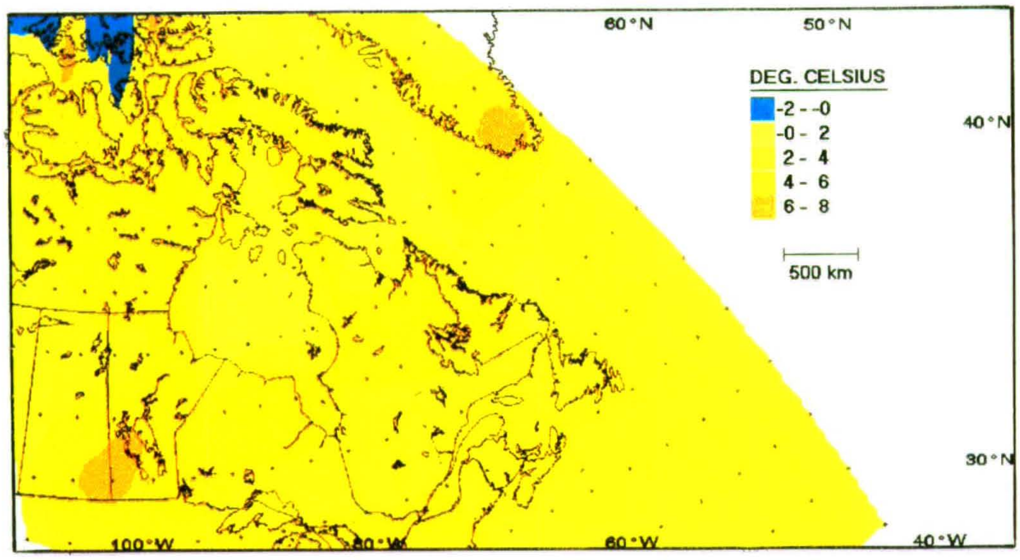
Source: Environment Canada

Appendix 3.4.2 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for spring daily maximum temperature.



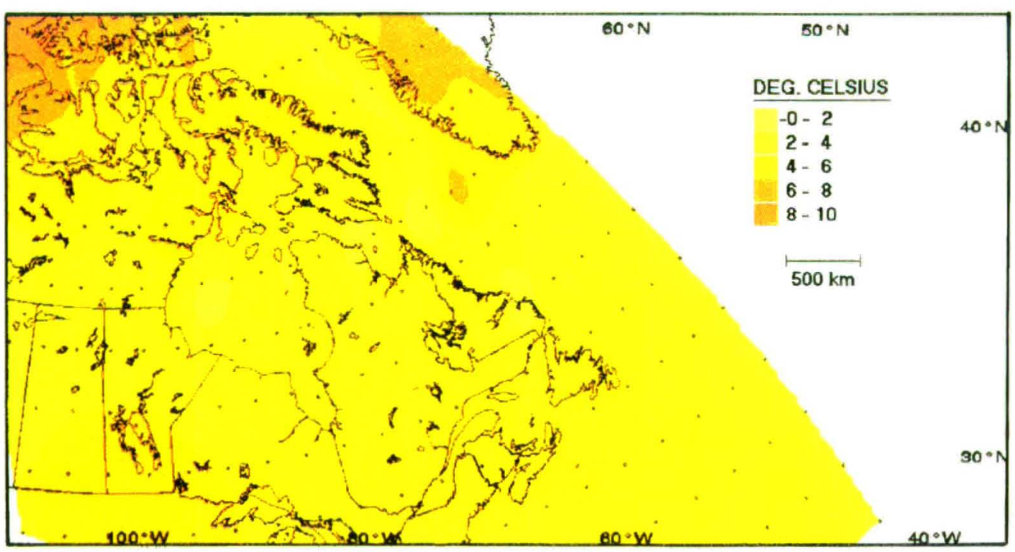
Source: Environment Canada

Appendix 3.4.3 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for summer daily maximum temperature.



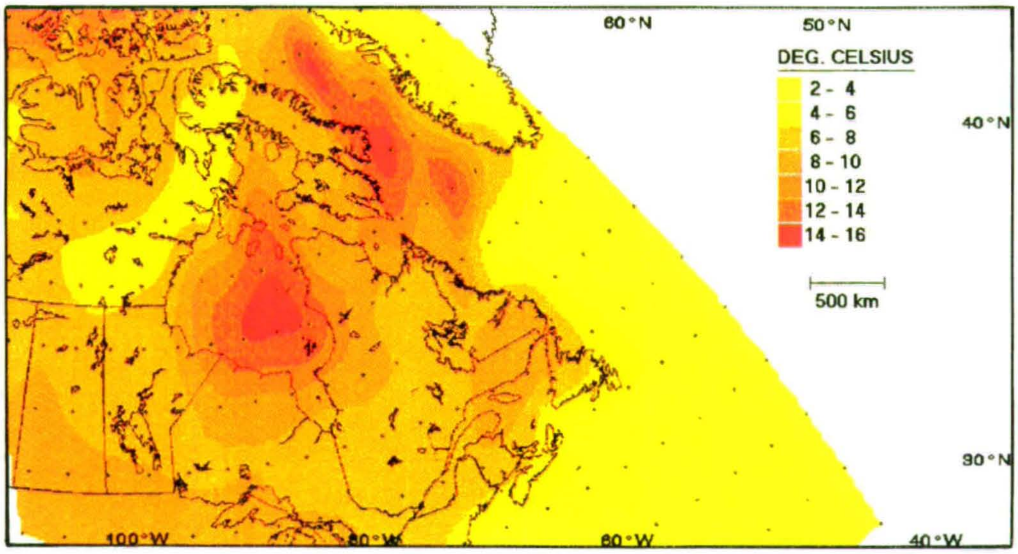
Source: Environment Canada

Appendix 3.4.4 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for autumn daily maximum temperature.



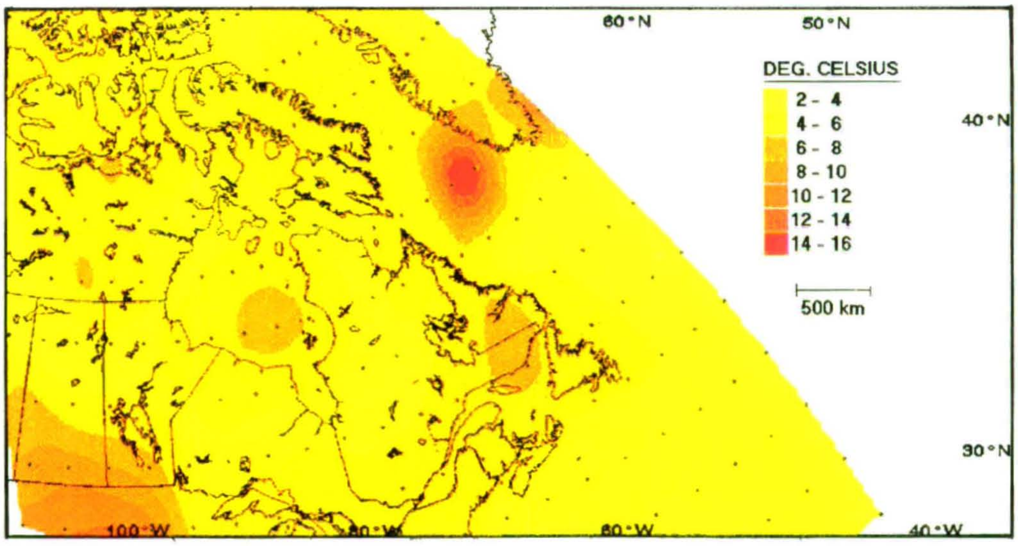
Source: Environment Canada

Appendix 3.4.5 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for winter daily minimum temperature.



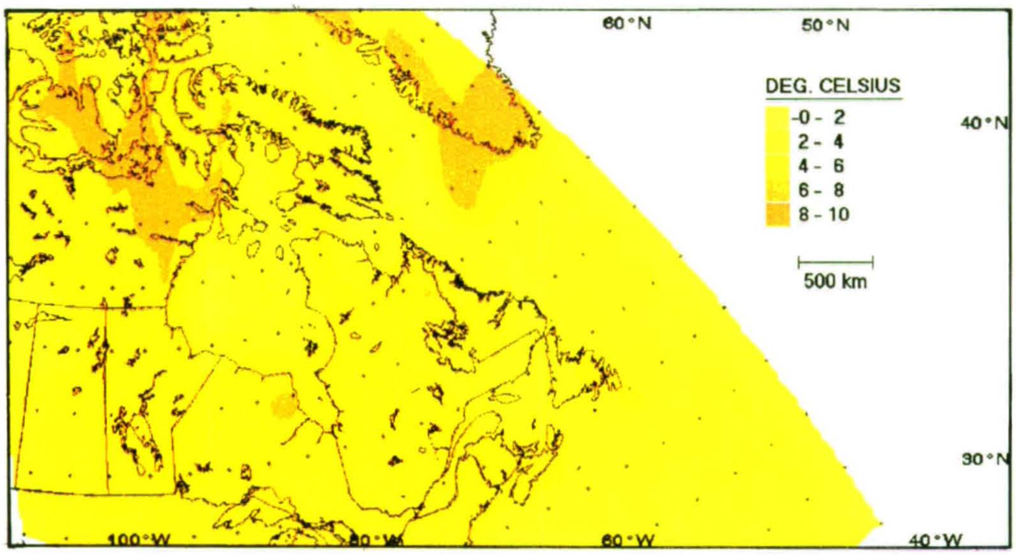
Source: Environment Canada

Appendix 3.4.6 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for spring daily minimum temperature.



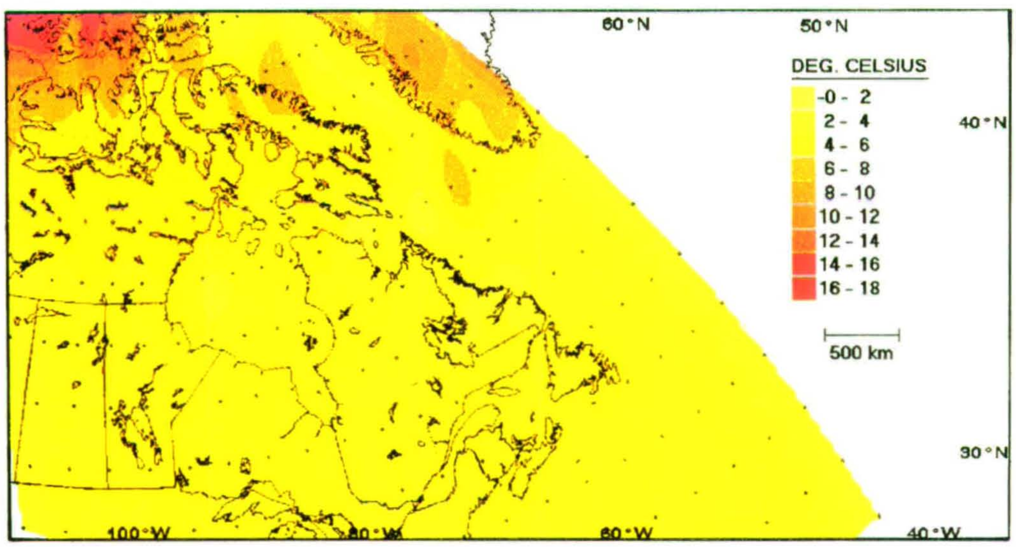
Source: Environment Canada

Appendix 3.4.7 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for summer daily minimum temperature.



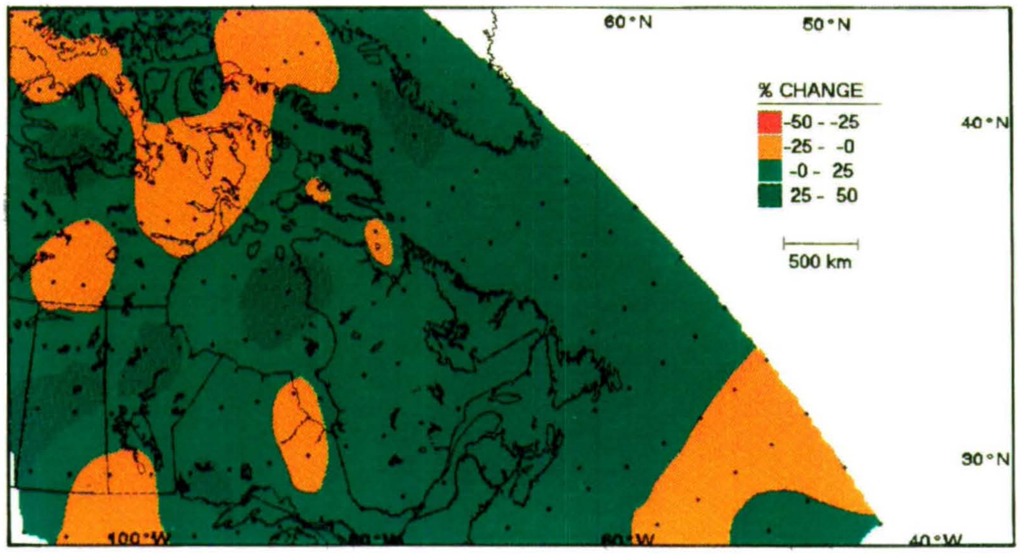
Source: Environment Canada

Appendix 3.4.8 CCC GCM ($2xCO_2$ - $1xCO_2$) scenario for autumn daily minimum temperature.



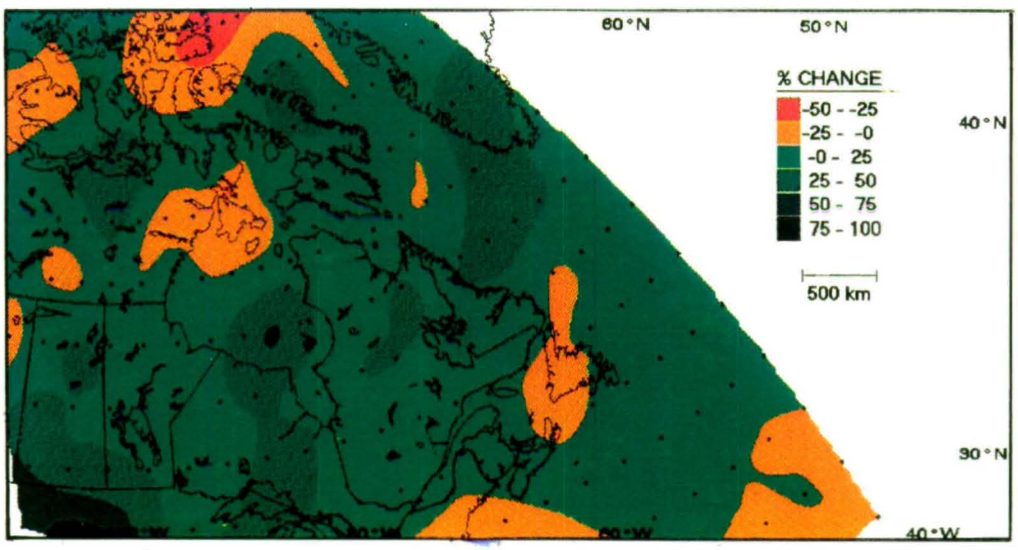
Source: Environment Canada

Appendix 3.4.9 CCC GCM total winter precipitation change scenario from $1xCO_2$ to $2xCO_2$.



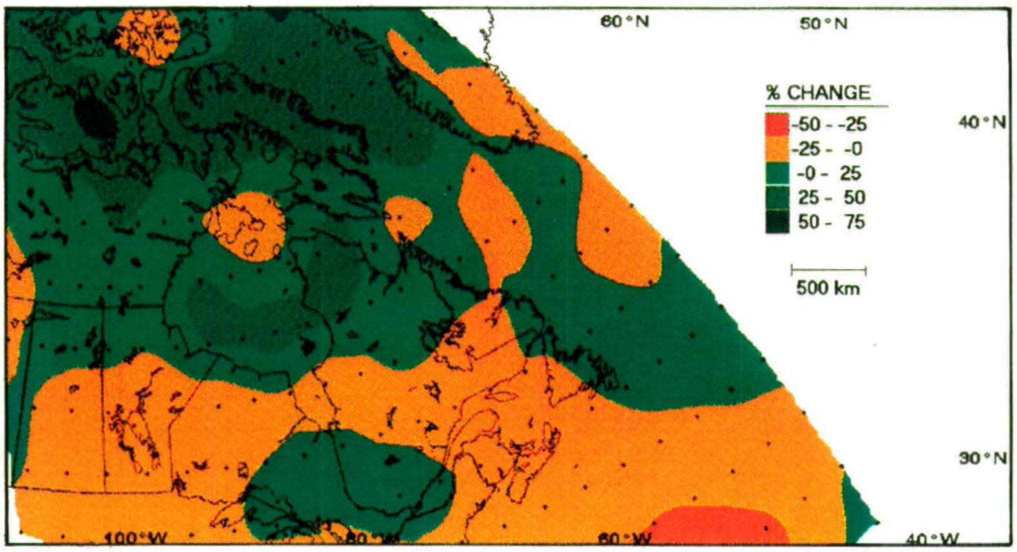
Source: Environment Canada

Appendix 3.4.10 CCC GCM total spring precipitation change scenario from $1xCO_2$ to $2xCO_2$



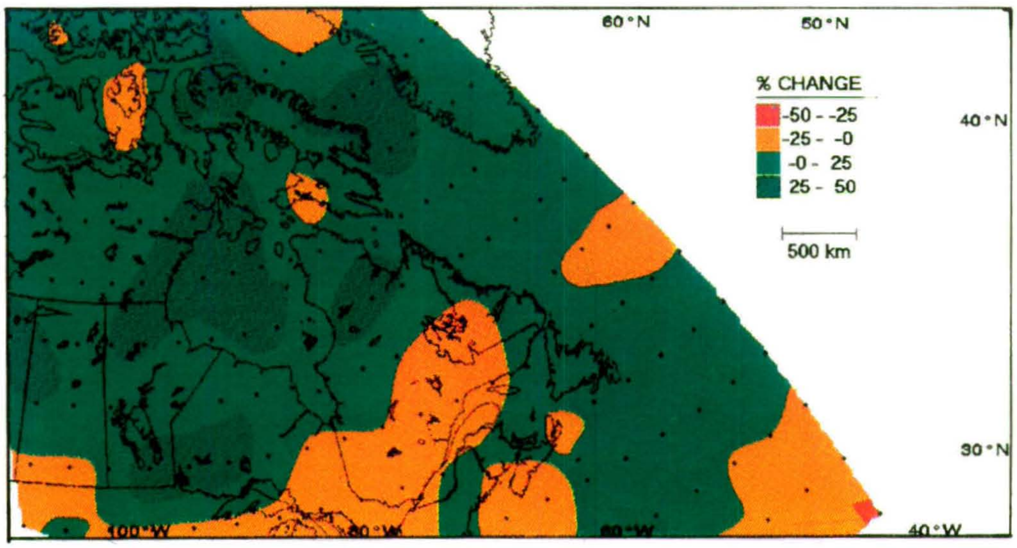
Source: Environment Canada

Appendix 3.4.11 CCC GCM total summer precipitation change scenario from $1xCO_2$ to $2xCO_2$.



Source: Environment Canada

Appendix 3.4.12 CCC GCM total autumn precipitation change scenario from $1xCO_2$ to $2xCO_2$



Source: Environment Canada