Canadian Cryospheric Response to an Anomalous Warm Summer: A Synthesis of the Climate Change Action Fund Project "The State of the Arctic Cryosphere during the Extreme Warm Summer of 1998"

D.E. Atkinson[1*], R. Brown[2], B. Alt[3], T. Agnew[4], J. Bourgeois[5], M. Burgess[5], C. Duguay[6†], G. Henry[7], S. Jeffers[8], R. Koerner[5], A.G. Lewkowicz[9], S. McCourt[8], H. Melling[10], M. Sharp[11], S. Smith[5], A. Walker[4], K. Wilson[8], S. Wolfe[5], M-k. Woo[12] and K.L. Young[13]

[1]Geological Survey of Canada - Atlantic, Natural Resources Canada, Dartmouth, Nova Scotia
[2]Meteorological Service of Canada, Environment Canada, Montréal, Quebec
[3]Balanced Environments Associates, Ottawa, Ontario
[4]Environment Canada, Toronto, Ontario
[5]Geological Survey of Canada, Natural Resources Canada, Ottawa, Ontario
[6]Université Laval, Québec, Quebec
[7]University of British Columbia, Vancouver, British Columbia
[8]Meteorological Service of Canada, Environment Canada, Ottawa, Ontario
[9]University of Ottawa, Ottawa, Ontario
[10]Institute of Ocean Sciences, Fisheries and Oceans Canada, Sidney, British Columbia
[11]University of Alberta, Edmonton, Alberta
[12]McMaster University, Hamilton, Ontario
[13]York University, Toronto, Ontario

[Original manuscript received 15 June 2004; in revised form 3 April 2006]

ABSTRACT *As of 2003, the warmest year on record in Canada (and globally) was 1998. Extensive warming was observed over the Canadian Arctic during the summer of 1998. A collaborative, interdisciplinary project involving government, universities, and the private sector examined the effect of this unusual warmth on cryospheric conditions and documented the responses, placing them in a 30–40 year context. This paper represents a synthesis of these results. 1998 was characterized by a melt season of exceptional length, having both an unusually early start and late finish. Extremes were noted for cryospheric variables that included ground thaw penetration, snow-free season, lake-ice-free season, glacier melt, and the duration and extent of marine open water. The warm conditions contributed to the break-up of two long-term, multi-year ice plugs in the north-west Canadian Arctic Archipelago, which allowed floe ice into the Northwest Passage. Synoptic events and preconditioning were observed to play an important role in governing the response of some variables to the warming. It was also noted that response was not uniform in all regions. This study provided an opportunity to examine possible cryospheric response to future, warmer conditions. It also provided a chance to assess the capability of current cryospheric monitoring networks in the Canadian Arctic. This study has suggested the manner of cryospheric response to low frequency, high magnitude events occurring within the broader milieu of large-scale forcing.*

RÉSUMÉ *[Traduit par la rédaction] En 2003, l'année la plus chaude au Canada (et dans le monde) était 1998. Un réchauffement généralisé a été observé dans l'Arctique canadien durant l'été 1998. Un projet coopératif interdisciplinaire, impliquant le gouvernement, différentes universités et le secteur privé, s'est intéressé à l'effet de ce temps chaud inhabituel sur les conditions de la cryosphère et a documenté les réponses en les plaçant dans un contexte de 30 à 40 ans. Le présent article constitue une synthèse de ces résultats. L'année 1998 a été caractérisée par une saison de fonte exceptionnellement longue, son commencement et sa fin s'étant produits nettement plus tôt et plus tard qu'à l'habitude, respectivement. On a observé des valeurs extrêmes pour les variables cryosphériques, notamment la profondeur du dégel dans le sol, la saison libre de neige, la saison libre de glace de lac, la fonte des glaciers et la durée et l'étendue des eaux libres marines. Le temps chaud a contribué à faire disparaître deux bouchons de glace de plusieurs années depuis longtemps présents dans le nord-ouest de l'archipel Arctique canadien, ne laissant que de la glace de dérive dans le passage du Nord-Ouest. On a observé*

*Corresponding author's e-mail: datkinson@iarc.uaf.edu; current affiliation - International Arctic Research Center, University of Alaska Fairbanks, 930 Koyukuk Drive, Fairbanks AK 99775-7335
†Current Affiliation - Geophysical Institute, University of Alaska Fairbanks, Fairbanks AK

que des phénomènes synoptiques et un préconditionnement ont joué un rôle important en orientant la réponse de certaines variables au réchauffement. On a aussi remarqué que la réponse n'était pas uniforme dans toutes les régions. Cette étude a fourni une occasion d'examiner la réponse cryosphérique possible à des conditions futures plus chaudes. Elle a aussi permis d'évaluer la capacité des réseaux actuels d'observation de la cryosphère dans l'Arctique canadien. L'étude a donné un aperçu du mode de réponse de la cryosphère à des évènements de faible fréquence et de grande amplitude se produisant dans le contexte plus général du forçage à grande échelle.

1 Introduction

The cryosphere, which includes sea ice, lake ice, river ice, snow cover, permafrost and seasonally frozen ground, ice caps and ice sheets, and glaciers, is an integral part of the global climate system with important linkages and feedbacks (Goodison et al., 1999). Recent studies consistently indicate extensive high latitude warming in the northern hemisphere over the last 50 years (Vinnikov et al., 1999; Magnuson et al., 2000; Serreze et al., 2000; Jones and Mohberg, 2003; Lanzante et al., 2003), with corresponding reductions in most components of the cryosphere (e.g., Comiso, 2002; Rouse et al., 2003a). Many of the warmest years occurred during the 1980s and 1990s. In the Canadian Arctic (Fig. 1), one year in particular, 1998, was conspicuous for the unprecedented warmth and duration of melting conditions over wide areas, affecting numerous variables (Tables 1 and 2; Fig. 2). This extreme event generated considerable interest in the Canadian scientific community, and a collaborative effort, "The State of the Arctic Cryosphere During the Extreme Warm Summer of 1998" project, supported by the Climate Change Action Fund (CCAF), was initiated in 2000–01 to determine how the major components of the Canadian Arctic cryosphere responded to this anomalous year, the spatial extent of the responses, and whether these responses lay outside observed variability. The CCAF endeavour, completed in 2005, was operated by a Canadian federal government ministry, Environment Canada.

This paper synthesizes results from a series of individual projects and reports prepared for the CCAF study. It includes a review of the evolution of the temperature anomaly and the associated atmospheric circulation patterns leading up to and during the summer of 1998, the response of various cryospheric components, discussion of important observations, and conclusions (CCAF Summer 1998 Project Team, 2001)[1]. Material presented consists of results from work performed specifically for this project, from work ongoing at the time, and previously available background information. Attempts to categorize the various datasets in terms of whether or not work was performed specifically for this project were not undertaken because the objective was a synthesis of material available at the time. The general arrangement is as follows: Section 2 details the antecedent and evolving climatic driving conditions, including the possible impact of the strong El Niño of 1997–98; Section 3 describes the response of several major components of the cryospheric system; Section 4 integrates these observations to arrive at a generalized cryospheric response; Section 5 discusses whether these observations are applicable analogues for hypothesized future "warm climate" scenarios; Section 6 ends with conclusions and recommendations. In general, if one of the authors of this paper has a post-1999 citation in this paper, that work will have contributed to this project.

2 Atmospheric conditions evolution

a *Temperature Anomaly Context and General Drivers*

1998 was one of the three warmest summers in the 51-year record across the majority of the Canadian north and alpine regions (with the exception of the Yukon and northern British Columbia mountains). The central and west-central Canadian Arctic Archipelago (CAA) and the Mackenzie District experienced their warmest summer, +2.5°C and +3.0°C above the 1951–80 normal, respectively, and their warmest annual temperatures on record. Comparison of the warm temperature anomaly of 1998 with the following four years shows that it remains one of the largest anomalies observed in this area (Figs 3 and 4). Comparison of the spatial pattern of the Canadian 1998 temperature anomaly with spatial patterns associated with temperature trends in Canada (Zhang et al., 2000) reveals a similarity between winter trends, but not summer trends, and the 1998 pattern (Fig. 5). Zhang et al. (2000) also indicate that a warming trend is not strong in summer. Cloud cover, an important driver of surface air temperature response, also shows little trend (Milewska, 2004), although cloud data are very difficult to recover from ocean areas such as the Beaufort Sea (Schweiger, 2004). These points suggest that the 1998 anomaly is not related to broad temperature trends. However, there is a strong correlation between winter (DJF) 500-hPa geopotential height and surface air temperature (Fig. 6) in the Yukon-Beaufort Sea-western archipelago region. This suggests a link between pressure and temperature response that was also strongly apparent in 1998, discussed below. Thus, although the spatial pattern associated with the 1998 warm anomaly does not match that observed for summer temperature trends, suggesting it to be an occurrence distinct from broader trends, the driver for the 1998 anomaly appears to be the same mechanism acting in the same region as that driving winter warming trends, namely, a pressure height anomaly over north-western Canada.

[1]The individual project reports that contributed to this synthesis (along with considerable additional material and data) are available on-line at the Canadian Cryopsheric Information Network (CCIN, www.ccin.ca). Where these have subsequently been published references are given.

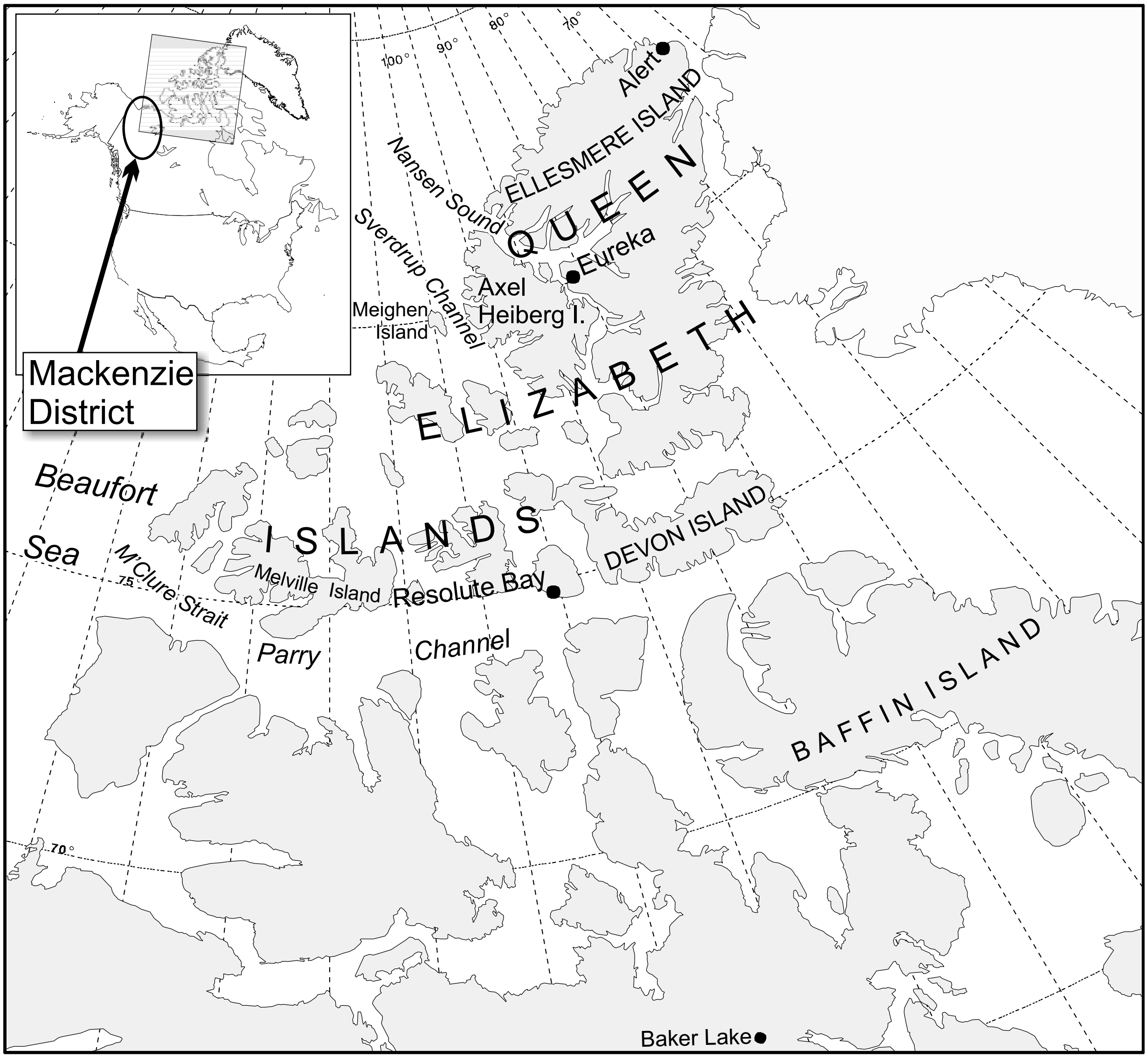

Fig. 1 Location map with the Canadian Arctic Archipelago (CAA) depicted. "Mackenzie District" and "Mackenzie Delta" are situated within the oval placed on the inset map.

b *Temperature Anomaly Evolution*

The evolution of the summer anomaly began in the winter of 1997–98 (not shown), during which below-normal temperatures occurred in the northern CAA and north-eastern Canada and above-normal temperatures occurred in south central Canada, with a secondary warm anomaly in the Yukon. By the spring of 1998 (not shown), a deep warm anomaly (up to +9ºC) stretched from the Mackenzie Delta and western Arctic Islands, covering all of the Mackenzie District, and reaching as far south as the southern British Columbia mountains. A weak, negative anomaly region was, by that point, limited to a small area of southern Baffin Island. By summer all northern and alpine regions of Canada were exhibiting above-normal temperatures, with the positive anomalies centred over the Mackenzie District, western Arctic Islands, Hudson Bay and the southern British Columbia mountains. In autumn the anomaly again intensified and shifted west to cover the CAA, extending into the Mackenzie District and Hudson Bay. In general, therefore, the warm anomaly for the May to October period (Fig. 3 – 1998) was centred over the Beaufort Sea and southern CAA. This strong and persistent warm anomaly generated an exceptionally long melt season in these areas (Fig. 7). "Melt season" here is defined as the period during which the 10-day mean 2 m air temperature is above zero. Thus the melt season for 1998 was, for many areas, both unusually warm and unusually long. Precipitation anomalies were also noted. Snowfall during the preceding winter (1997–98) was characterized by below-average precipitation over the western and High Arctic, and above-average precipitation over western Hudson Bay and the eastern Arctic (Fig. 8). Both of these aspects will be discussed further in the text.

TABLE 1. Extremes in 1998 Arctic climate and cryospheric components

- Highest annual temperatures on record both in the west (Tuktoyaktuk, 1956–98) and the High (Resolute, 1950–98) Arctic
- Second highest average summer temperatures (highest–1962) in QEI (1950–98)
- Warmest summer in the Mackenzie and Arctic Tundra region (1948–98)
- Second warmest summer (warmest – 1981) in the Arctic Mountains and fiords region (1948–98)
- Earliest snow melt in the western Arctic (1955–98)
- Latest start to the snow season in the eastern Arctic (1955–98)
- Record minimum ice extent in the CAA (1961–2000)
- Latest date for maximum open water QEI (1961–2000)
- Record (1958–98) northward retreat of the ice edge in the Beaufort Sea west of 141°W longitude
- Second most negative mass balance on the Devon Ice Cap QEI (1961–99)
- Active layer detachment slide activity in both the west and High Arctic
- Extension of thaw ground season, 8–25 days longer than previous years, in the MkD (1991–99)
- 50–90% greater ground thaw degree index than in previous years in the MkD (1991–99)
- 12–23 cm greater thaw penetration in the MkD (1991–99)
- Earliest observed ice break-up dates on Great Slave and Great Bear lakes (1988–99)
- Latest observed freeze-up on Great Slave and Great Bear lakes (1988–98)
- Second thinnest maximum lake ice on Upper Dumbell Lake QEI (1960–98)

Location key:
MkD= Mackenzie Delta, E=eastern Canadian Arctic, W=western Canadian Arctic, QEI=Queen Elizabeth Islands, CAA=Canadian Arctic Archipelago

c *Atmospheric Circulation Anomalies*

The development of the anomalous melt season regime may be traced to the atmospheric circulation patterns in place at the time. There were several departures from mean circulation patterns during 1998. The primary departure consisted of a strong pressure ridge, present from the surface up to 500 hPa, over the CAA and extending to Greenland during June, July and August (Fig. 9). The high pressure anomaly displaced the polar vortex to the Siberian sector of the Arctic Ocean and caused the Beaufort Sea High to shift to just north of the CAA. During the winter and spring of 1998, the Aleutian Low also intensified and shifted to the north and west. This trough-ridge pattern established a strengthened pressure gradient across the region, such that a strong, southerly flow resulted over the Canadian Arctic (considered herein to include the CAA, north mainland coast and continental interior north of 60°N, and the Beaufort Sea). Subsidence due to the ridge over the CAA also occurred, favouring reduced cloudiness (Table 3). The result was above-average temperatures over the Canadian Arctic for an extended duration from the late winter to early spring of 1997–98 and most of 1998. The high pressure anomaly persisted well into September (Figs 10a and 10b), strengthening the weak ridge that is usually building by then. This was an especially unusual event over the Queen Elizabeth Islands (QEI) and eastern Arctic (Fig 10c), resulting in continued southerly flow and subsidence. At that time, 1998 was the warmest year in the instrumental record for the Canadian Arctic (since ~1950 for the High Arctic climate station network), and was also the warmest in the instrumental record for the northern hemisphere land area (since ~1850). Comparison with more recent

TABLE 2. Summary of the spatial characteristics of the response of the Canadian Arctic cryosphere to the extreme warming during the summer of 1998.

Cryospheric Variable	Spatial Pattern
Snow	The 1998 melt season was characterized by an initial early retreat of the spring snow-line over the entire North American continent in April and May. Early spring warming coupled with below-average winter snow accumulation over western Canada and the Mackenzie Basin, led to record (1955–2000) early melt of snow across the western Arctic. Spring snow cover also retreated early across the eastern Arctic (second earliest during the 1955–2000 period, earliest occurring in 1994), but the anomaly was more marked during the fall across the eastern Arctic where the prolonged warmth led to a record (1955–2000) late start to the snow season. Snowfall occurred in late July in northern Ellesmere Island, which did not experience the prolonged summer warming observed further south.
Sea Ice	In the west, break-up was early and there was twice as much open water in the southern Beaufort sea as normal; distance to the ice edge from the Alaskan coast was 46% greater than for the previous record in 1954. Minimum ice extent conditions extended into the QEI where the latest date for minimum ice extent during the previous 40 years was recorded. North-western Baffin Bay and the south-eastern QEI channels were open compared to other years during the 1990s but not compared to the remainder of the 40-year record. Total accumulated coverage for the eastern Arctic and Hudson Bay was the least during the 30-year record. Break-up was early and freeze-up late in Hudson Bay.
Lake Ice	In the west, the earliest break-up and latest freeze-up in the twelve-year record was observed on Great Slave and Great Bear lakes and the ice free season was generally long across the north-western mainland. In the south-central Arctic (Baker Lake) freeze-up was relatively late. In the far north-east (Alert), maximum lake-ice thickness was the second thinnest in the 30-year record.
Glaciers and Ice Caps	In the west, the Melville Ice Cap experienced the most negative mass balance on record (but the record does not include the warm seasons of 1960 and 1962) and had both an early start and late finish to the melt season. In the south-eastern QEI, the Devon Ice Cap had a highly negative balance (surpassed only in the early 1960s) and at low elevations the melt season was prolonged. In the north (Meighen and Axel Heiberg islands) a late July snow storm brought an early end to the melt season and near normal mass balance conditions.
Permafrost and Active Layer	In the west, shallow permafrost temperatures and thaw penetration were generally the highest in the ten-year record and there was evidence of ground ice wedge melt and increased active layer detachment slide (ALDS) activity. In the south-central Arctic (Baker Lake) freezing of the active layer was delayed by warm fall temperatures. To the north, shallow ground temperatures at Resolute were the highest in the 30 years of record and increased ALDS activity was observed in the interior of Ellesmere Island.

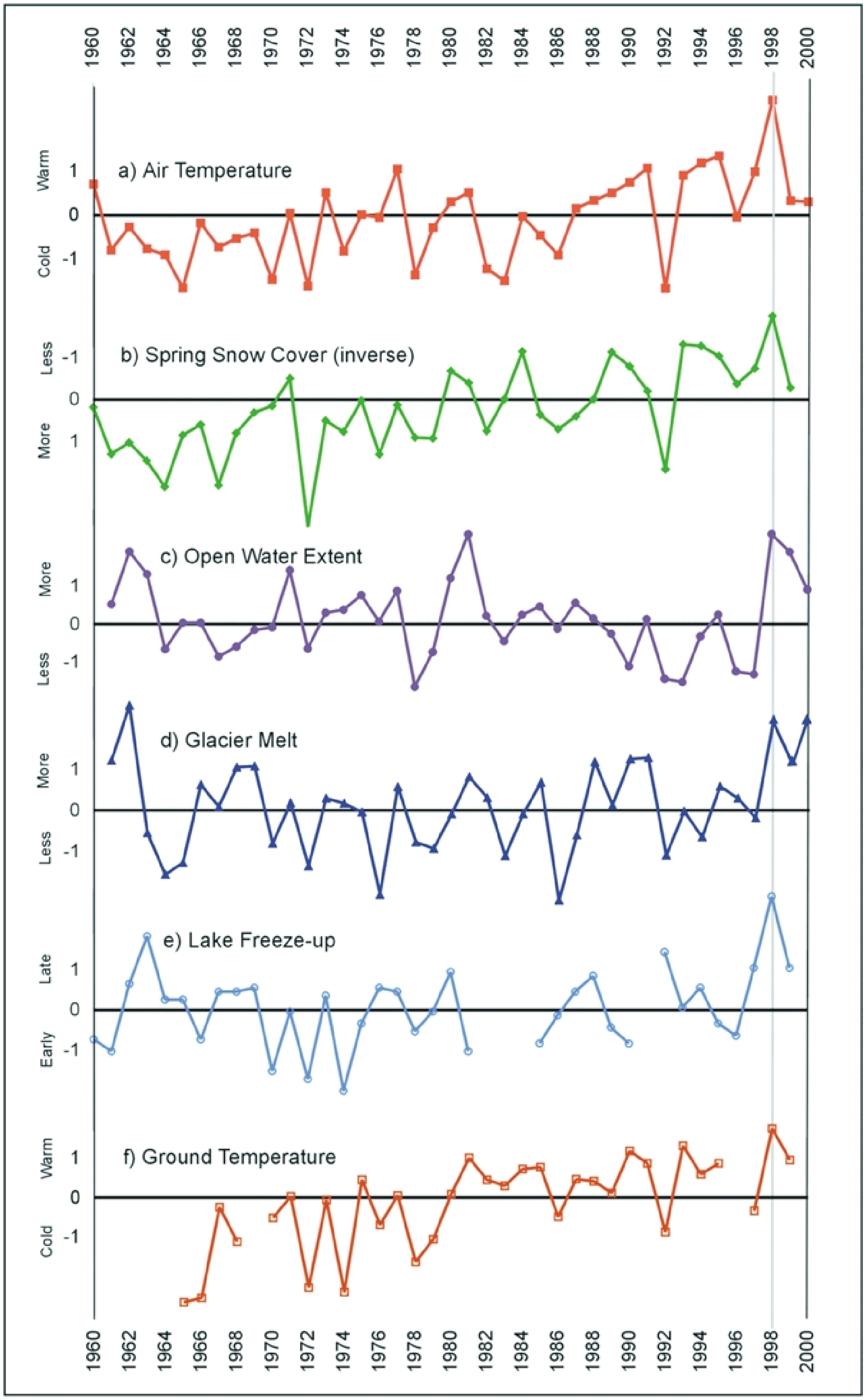

Fig. 2 Time series of selected cryospheric variables: (a) Canadian Arctic (north of 65°) May–October temperature anomalies for the 1960–2000 period; (b) spring snow cover duration at Canadian Arctic climate stations; (c) maximum open water extent in the QEI; (d) summer melt of the Devon Ice Cap; (e) date of lake freeze-up, Great Slave Lake; and (f) July 100 cm ground temperature, Resolute. All series are normalized with respect to 1968–98 (i.e., anomaly divided by the standard deviation). Details on the construction of these time series are provided in the Appendix. In general it is important to note that the observation of anomalous warmth in 1998 (or any year) represents mean conditions over a large area. The anomaly is not necessarily uniform across the entire CAA and Canadian Arctic, nor is it uniform over the entire melt season.

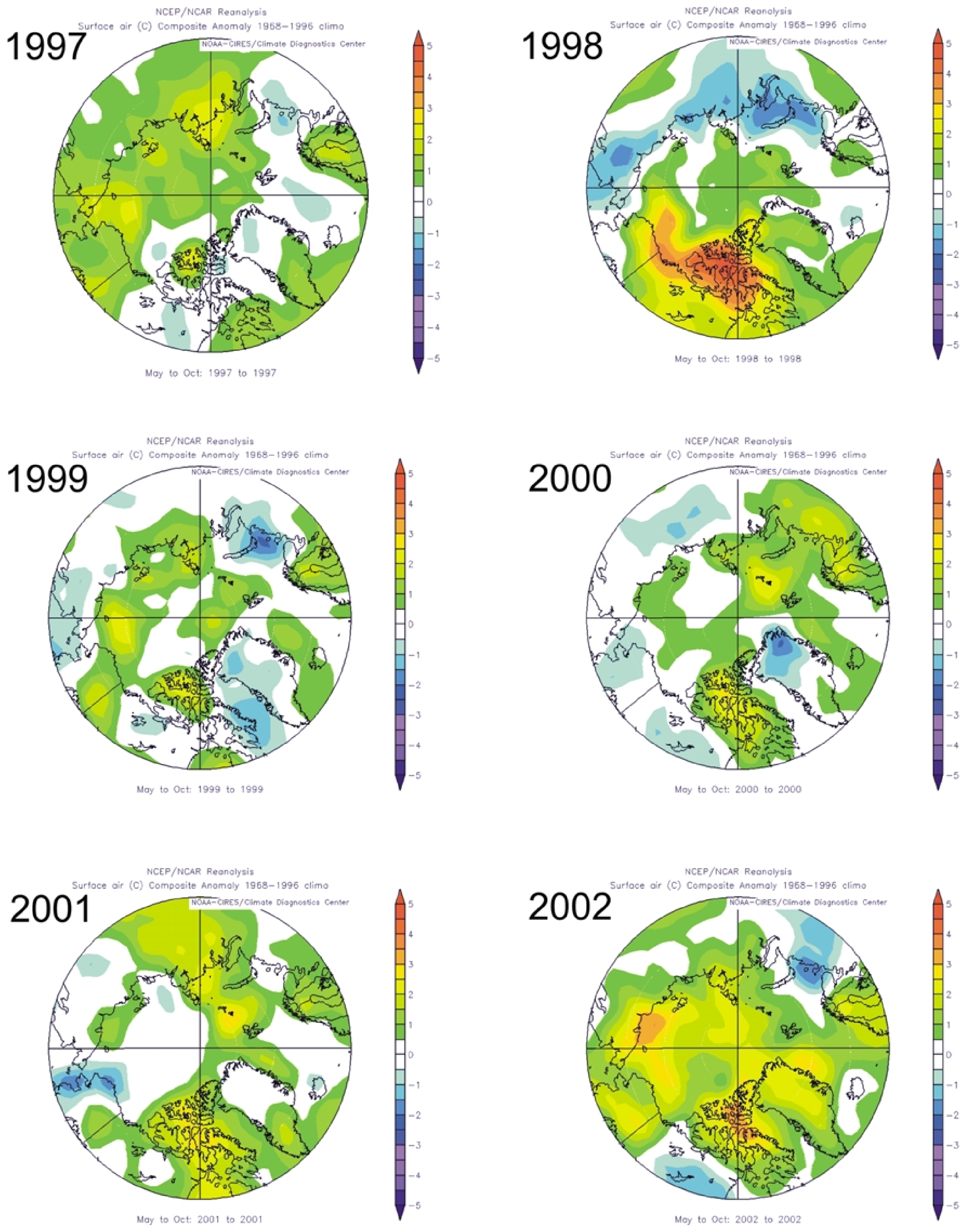

Fig. 3 Surface air temperature (2 m height) mean anomalies for the years 1997–2002 computed from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis monthly data; the colour scale is uniform for all plots. Note that local and regional departures from plotted values are possible. These come from reanalysis resolution shortcomings in representing some of the more topographically complex regions (NOAA-CIRES Climate Diagnostics Center, Boulder Colorado from their web site at http://www.cdc.noaa.gov/).

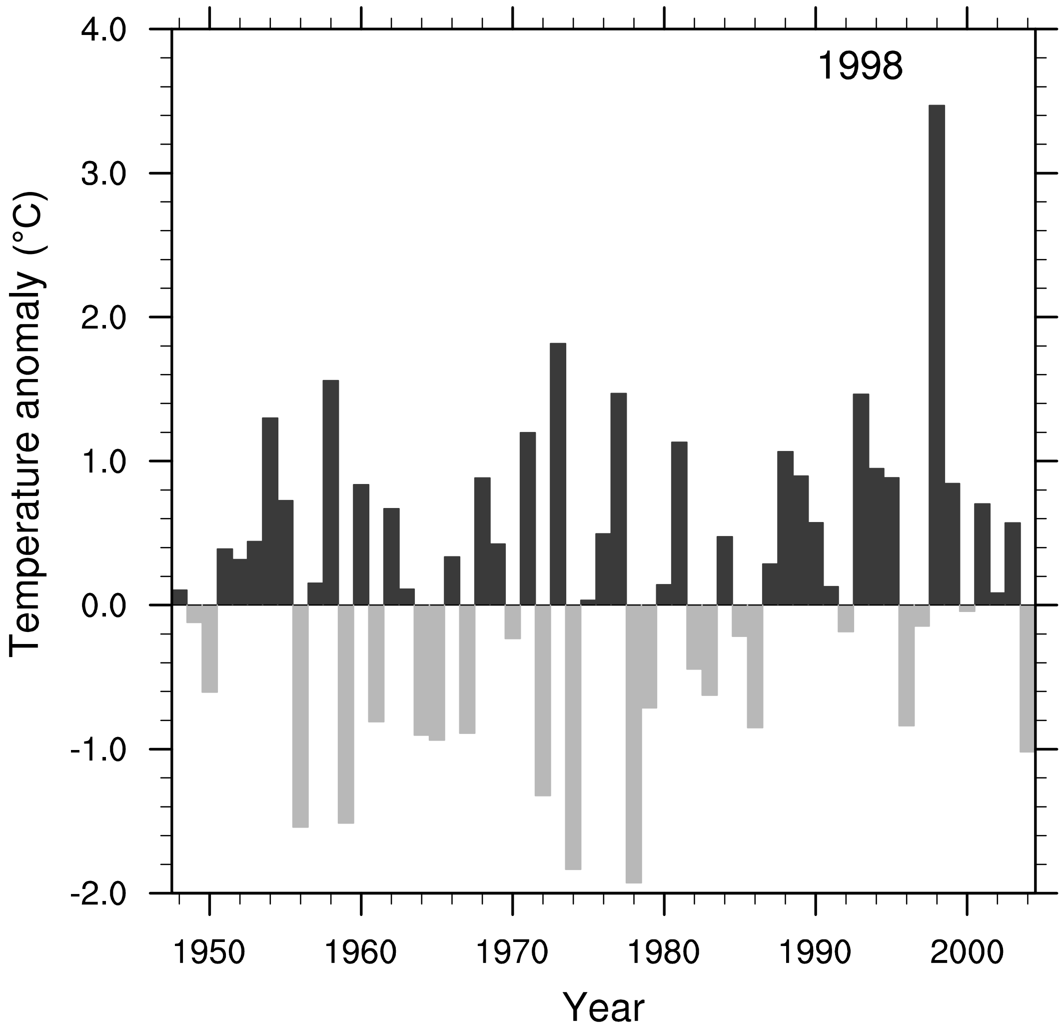

Fig. 4 Temperature means for May–October over 67.5°–82.5°N (4 grid points) and 102.5°–137.5°W (8 grid points) using the Jones observationally derived gridded dataset from the Climatic Research Unit, East Anglia (Jones and Moberg, 2003).

data shows that, while on average for the Canadian Arctic the four summers and autumns since 1998 (i.e., 1999 to 2002) were also warmer than the 1968–96 mean, 1998 remains the warmest on record even within a generally anomalous warm period (Fig. 3), especially for autumn.

d *Potential Connection to theEl Niño Southern Oscillation*

The possible influence of the El Niño Southern Oscillation (ENSO) in creating or aiding the melt season anomaly was also investigated because the 1997–98 ENSO event was one of the strongest of the twentieth century, and the observed circulation and temperature anomaly patterns are consistent with typical high-latitude responses to ENSO. This was also motivated by the fact that the US National Climatic Data Center (NCDC) concluded, "A persistent El Niño in the first half of the year and the unprecedented warmth of the Indian Ocean contributed to this record warm year [globally]." (www.ncdc.noaa.gov/oa/climate/research/1998/ann/ann98.html). A variety of researchers have noted apparent links between ENSO forcing and the response of the Arctic climate system (e.g., Ono, 1996; Gloersen, 1995). In the North American Arctic, these links appears to work most directly through the Aleutian Low (e.g., Niebauer, 1999). In fact, the Aleutian Low in late 1997 to early 1998 was both displaced to the north-west and unusually intense. It is also known that a deepening of the quasi-stationary pressure ridge over western Canada is linked to ENSO. Both of these circulation shifts will enhance southerly flow into the western Canadian Arctic. Given that this ENSO was strong through the winter and spring of 1998, it is possible that ENSO-related circulation changes could have contributed to the early start to the summer season. However, it is more difficult to argue that ENSO was linked to the extension of the summer season or to the Arctic storms that were responsible for the break-up of the ice plugs. This occurs, in part, because the Aleutian Low does not exert a strong influence on the climate of the CAA in the late summer; as well, the ENSO event of 1998 had weakened by mid-summer. It must also be noted that there is considerable variability in regional response to ENSO that arises from the variable nature of the ENSO events themselves, and from the

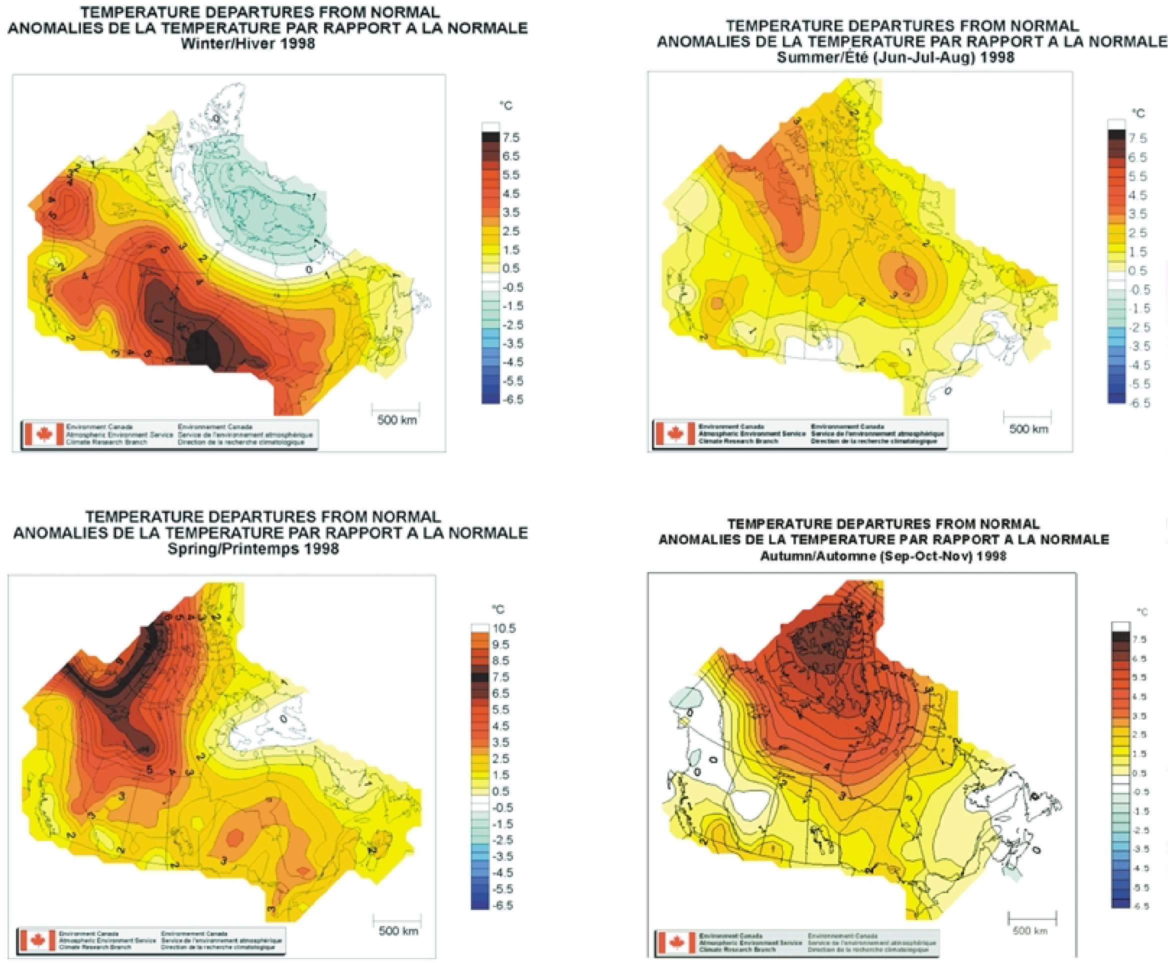

Fig. 5 Trends in surface air temperature (2 m height) maxima over Canada by season, from observational data. Maxima and minima were provided in Zhang et al. (2000); because their patterns were very similar only the figure for the maxima is reproduced. The 1998 anomaly pattern is similar to the pattern exhibited by the winter trend but not the summer trend (adapted with permission from Zhang et al. (2000)).

fact that, by inducing remote pressure anomalies, ENSO can be considered to be acting through more local pressure patterns, such as the Pacific-North America (PNA) Index or Western Pacific (WP) pattern. Given that they operate at different timescales (Papineau, 2001), they can act to enhance or damp the local expression of an ENSO signal, and so act to disrupt a clear correlation between ENSO and a high-latitude response. For example, ENSO years and sea-ice extremes in the Beaufort Sea and CAA do not necessarily coincide. More significantly, another year of extreme low sea-ice extent (1962) in the CAA did not coincide with a major ENSO event, and another very strong ENSO event in 1982 did not correspond to unusual warming in the Arctic. Despite this, given the known influence of ENSO on the western North America ridge and on the Aleutian Low (dipoles of the PNA pressure pattern oscillation), and their subsequent effects on the thermal regime of the western Canadian Arctic, it may be concluded that the strong ENSO event of 1997–98 played a dominant role in the climate of the western Canadian Arctic during the first half of 1998.

e *Comparison with Other Warm Years*

Inspection of the cryospheric time series (Fig. 2) revealed that other years have experienced cryospheric conditions similar to those of 1998. The most notable of these was 1962, during which there were strong anomalies (second largest on record) of open water extent, lake ice melt, and glacier melt, especially over the CAA. The summer of 1962 was also accompanied by a shift of the polar vortex to the Siberian sector of the Arctic Ocean (Alt, 1987; Agnew et al., 2001). In contrast to 1998, however, surface air temperatures (SAT) were below normal (Fig. 2). To provide a better assessment of temperature anomalies for these two summers, a detailed comparison of July temperatures (Fig. 11) was carried out using a regional, topo-climatic model (1 km resolution) driven with observed upper air data which took account of onshore coastal advection and elevation effects (Atkinson and Gajewski, 2002). The comparison showed that 1998 was warmer than 1962 over most of the Canadian Arctic (particularly in the west), with the exception of the QEI in the northern CAA and eastern Baffin Island. The cooler temperatures

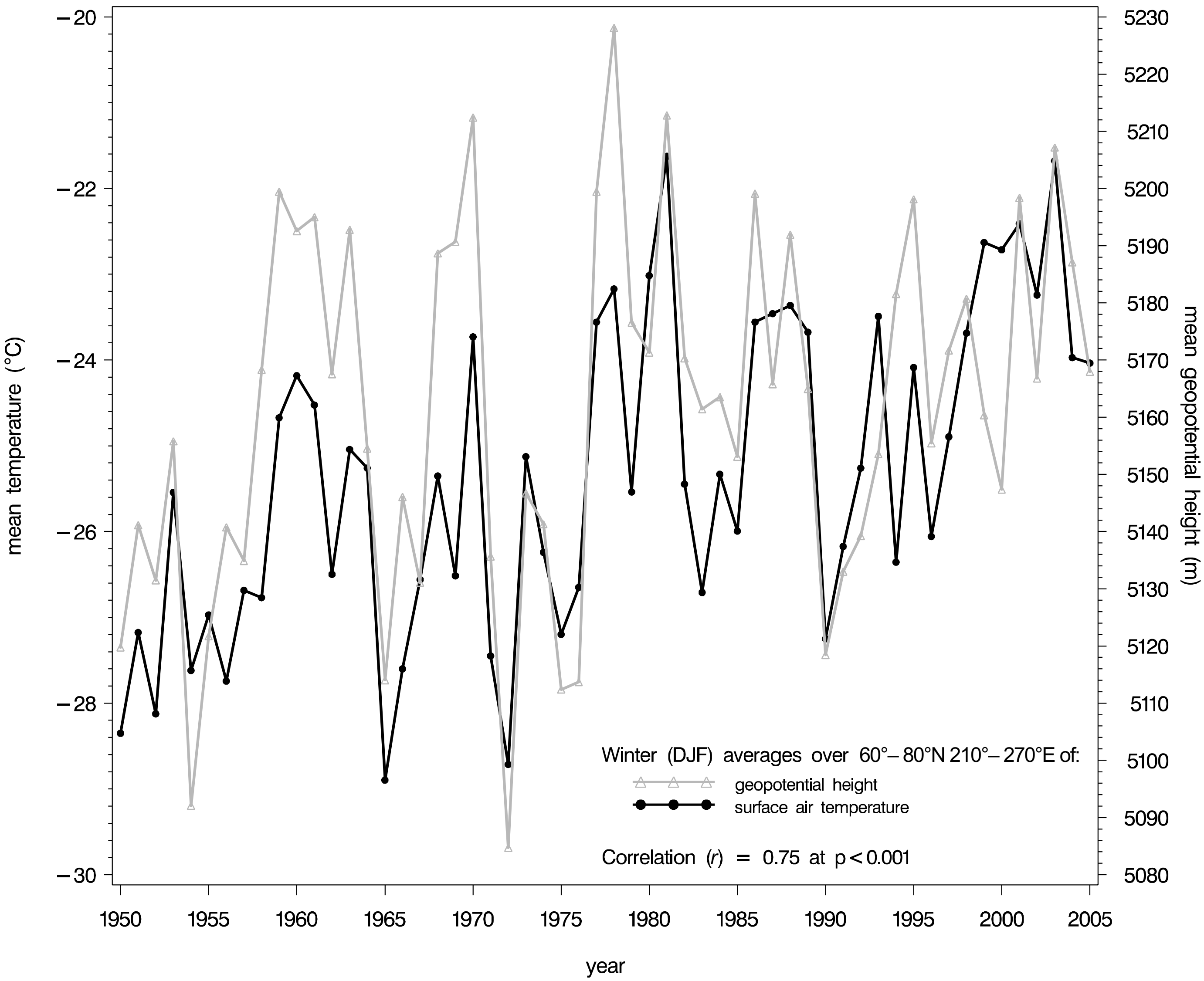

Fig. 6 Surface air temperature (°C) (black line) and 500 hPa geopotential height (m) (grey line) values from NCEP/NCAR reanalysis monthly data for 1950–2005. Data averaged over the grid box 60°–80°N and 210°–270°W for the climatological winter months (DJF). Correlation (r) = 0.75 ($p < 0.001$).

in the QEI are in agreement with observations from automated weather stations on eastern Ellesmere Island (not shown), which showed a smaller temperature anomaly than observed in the west during 1998.

3 Cryospheric response

a *Sea Ice*

A major focus of this project was to document sea-ice conditions in the Canadian Arctic and to examine interannual variability in ice cover and open water using historical records, satellite data, and weekly ice charts from the Canadian Ice Service (CIS). The weekly charts were required because the resolution of passive microwave satellite data (Scanning Multichannel Microwave Radiometer (SMMR) and Special Sensor Microwave Imager (SMM/I)) is too coarse to monitor sea-ice concentrations in this region. The interest in sea-ice, with respect to the CAA in particular, stems from concerns about shipping safety and from evidence that both multi-year ice plugs blocking the Sverdrup Channel and Nansen Sound cleared completely in 1998 (Fig. 12). This raised the question as to whether the absence of the plugs would allow enhanced penetration of Arctic Ocean multi-year ice into the CAA in subsequent years. RADARSAT imagery was used to document the break-up of the multi-year ice plugs as well as ice motion in the plug areas.

Anomalous warmth and a persistent south-easterly wind during the spring and summer of 1998 contributed to a marked reduction in sea-ice cover over the Beaufort Sea due to the melting and displacement of ice floes (Fig. 12a) and extensive clearing of ice from the QEI region of the CAA

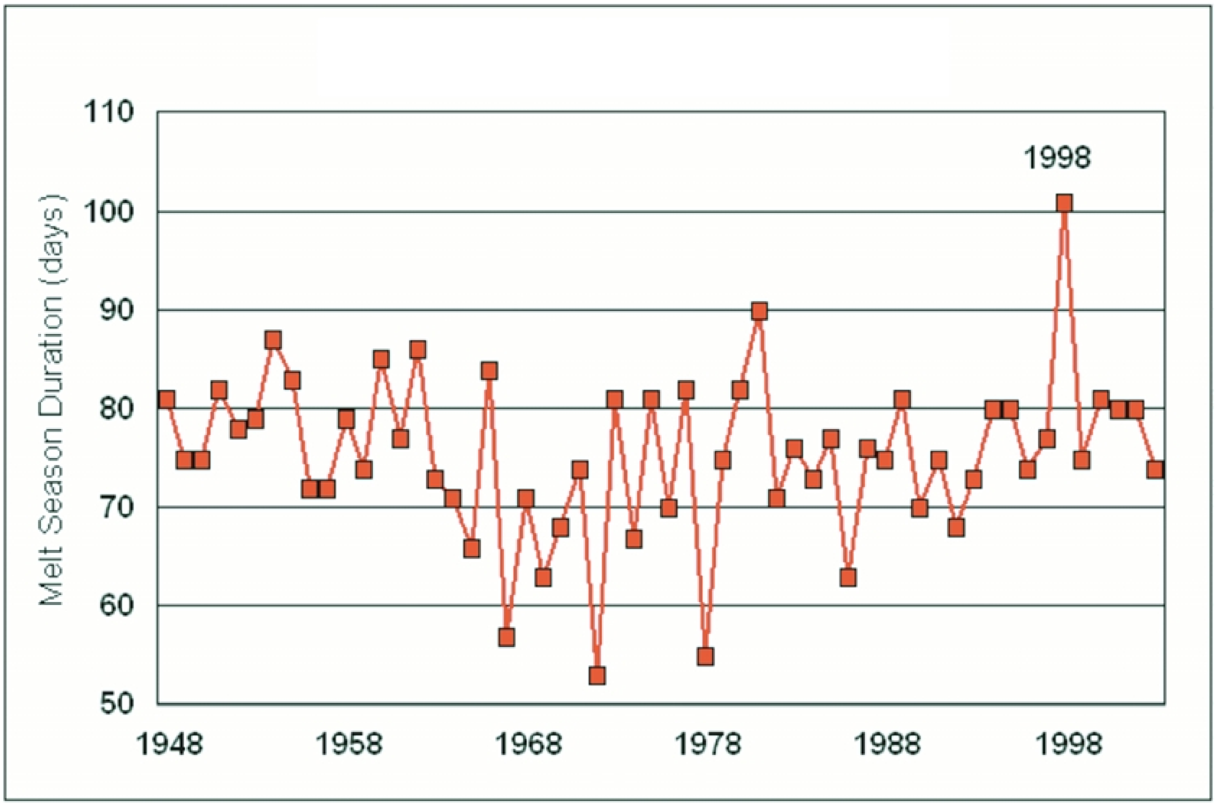

Fig. 7 Mean annual melt season lengths for the southern Canadian Arctic Archipelago, Beaufort Sea, and north continental mainland, derived from NCEP/NCAR reanalysis daily air temperature data (Kalnay et al., 1996).

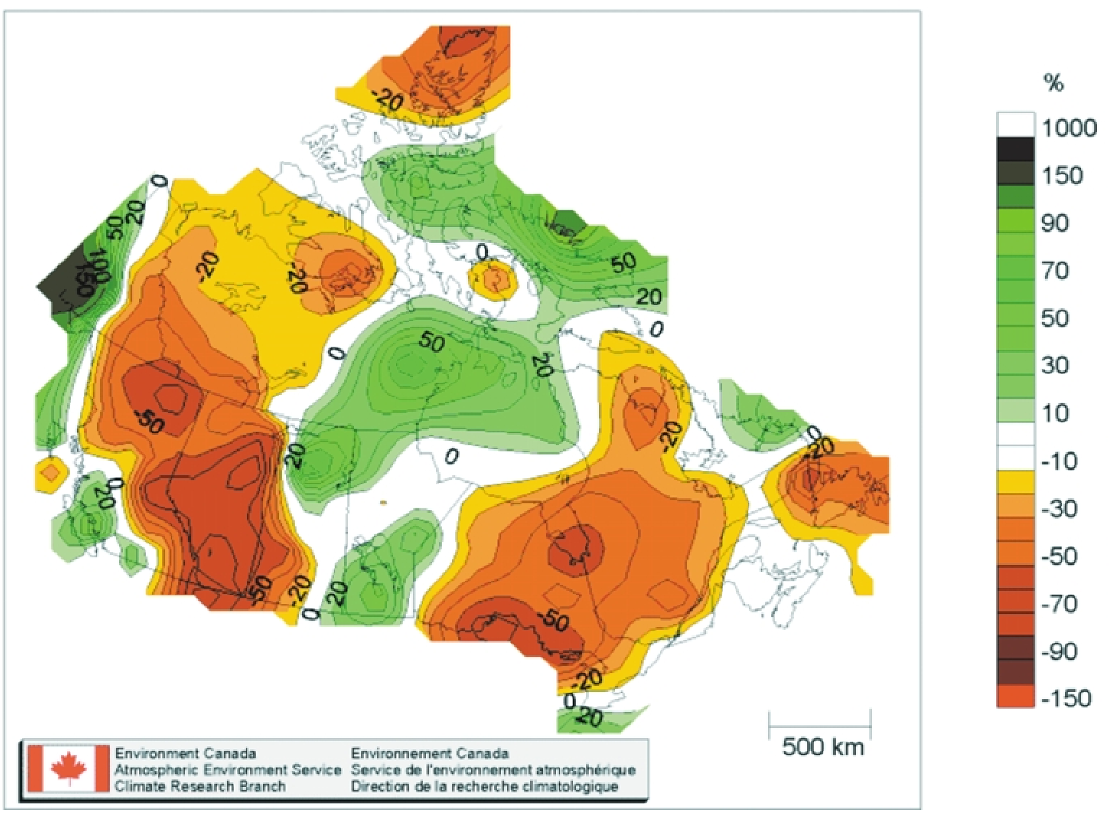

Fig. 8 Winter (DJF) precipitation anomaly for 1997–98 as a percentage of the 1951–80 mean. Positive values indicate precipitation totals exceeded the mean, negative values, were less than the mean (reproduced from Environment Canada, 2002).

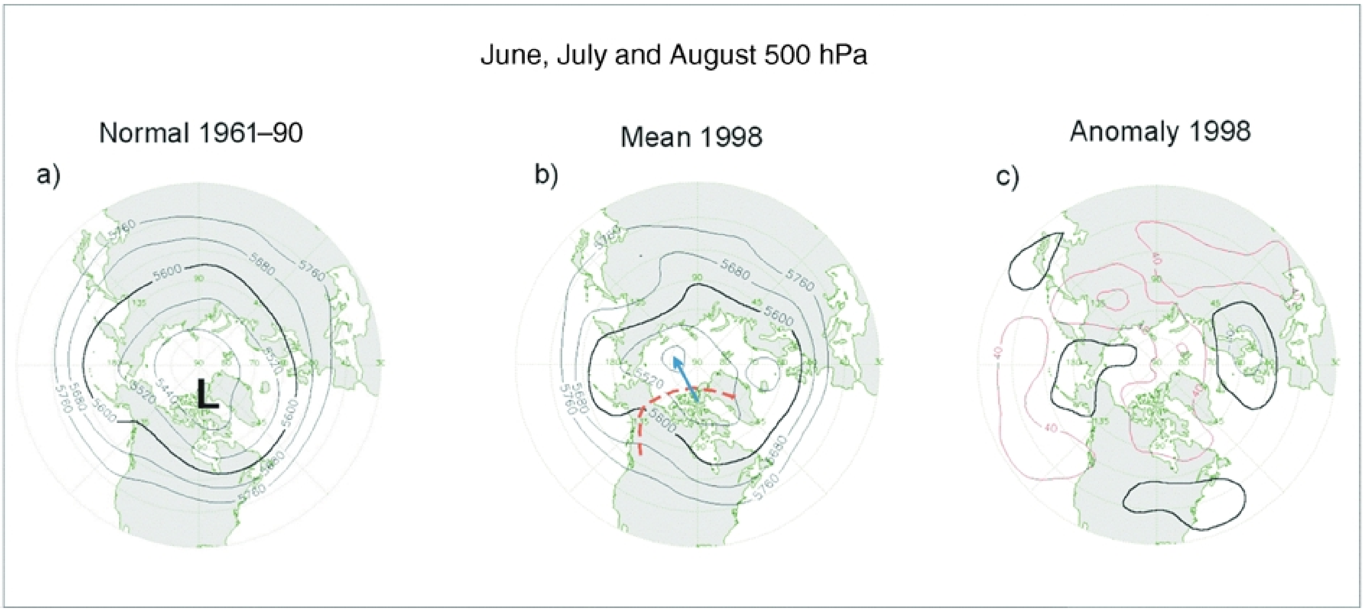

Fig. 9 Summer (June–August) 500 hPa heights: a) normal pattern for 1961–90 (deepest contour is 5440 m geopotential height), b) mean pattern for summer 1998 (small contour north of the East Siberian Sea is 5440 m geopotential height); and c) anomaly pattern for summer 1998 (red indicates the area of higher than normal heights; the line of zero variation is in bold). In b), the blue arrow indicates the displacement of the polar vortex centre from its mean position in the CAA (indicated by the L) to the East Siberian Sea sector, and the red dashed line indicates the 500 hPa ridge position.

TABLE 3. Comparison of cloud conditions with 1971–2000 normals by month for summer 1998, for Resolute Bay. Values indicate the percentage of observations for that month within the range category given under "Cloud cover amount".

	June		July		August	
Cloud cover amount (tenths)	normal (%)	1998 (%)	normal (%)	1998 (%)	normal (%)	1998 (%)
0–2	16	17	15	13	9	10
3–7	16	63	18	35	14	29
8–10	68	30	67	52	77	61

(Fig. 12b). The magnitude of the reduction in sea-ice extent over the western Arctic Ocean is clearly shown in the passive microwave record (Figs 13 and 14). Time series of the average latitude of ice extent in the Beaufort Sea for 1969–98 (not shown), and the distance to the ice edge north of Barrow, Alaska for 1954–98, confirmed that in 1998 the open-water area north of Alaska was the largest during the 1954–98 period (Maslanik et al., 1999). In the Canadian sector of the Beaufort Sea, the maximum ice-free area was comparable to other years, notably 1958, 1972, 1977, 1981, 1982, 1987, 1993 and 1995. Without the ice cover, insolation began warming the upper levels of the Beaufort Sea as early as late April 1998. By early August the sea temperature was 10°C to a 20 m depth over a wide area. In early September, warm water from this reservoir was observed flowing eastward into M'Clure Strait (Institute of Ocean Sciences cruise #1998-38). This inflow may have contributed to the very rapid loss of multi-year ice in western Parry Channel that occurred during August.

To examine the magnitude and variability of open water conditions in the CAA, a 40-year time series of minimum sea-ice extent (or open-water extent) was constructed from Polar Continental Shelf Project sea-ice charts (1961–78) and CIS digitized weekly ice charts (1968–2000). The homogenized series (Fig. 2c) showed two previous summers with comparable minimum sea-ice extent (1962 and 1981). However, the latest observed date of minimum sea-ice extent (in the 40-year period of record was 28 September 1998), and only in 1962 and 1998 did both the Sverdrup and Nansen ice plugs break up (Jeffers et al., 2001). The absence of a trend in the QEI minimum sea-ice extent series (Agnew et al., 2001; Jeffers et al., 2001) contrasts with the decreasing trend observed over the Arctic Ocean (Johannessen et al., 1999; Parkinson et al., 1999). However, this regional difference is consistent with studies documenting asymmetric regional trends in sea ice (Parkinson, 1992; Maslanik et al., 1996). Work by Proshutinsky et al. (2002) on the Beaufort Gyre suggests a possible mechanism for differential regional controls. The study of distinctive ice regimes within the QEI showed a slight (not statistically significant) trend towards less ice in the north and west and a statistically significant trend towards more ice in the south-east during the 1961–98 period (Fig. 15). The latter result is in agreement with temperature trends (Chapman and Walsh, 1993; Zhang et al., 2000) and paleoclimatic reconstructions of sea-ice extent (Grumet et al., 2001).

A detailed investigation of the break-up of the Nansen and Sverdrup ice plugs in 1998 was made possible by using RADARSAT data. According to Jefferies et al. (1992), the plugs are not an ice shelf or single homogenous blocks of ice; rather they are "an agglomeration of ice of different ages and thickness". In normal years, these plugs remain intact during the summer and thus their break-up indicates extreme conditions

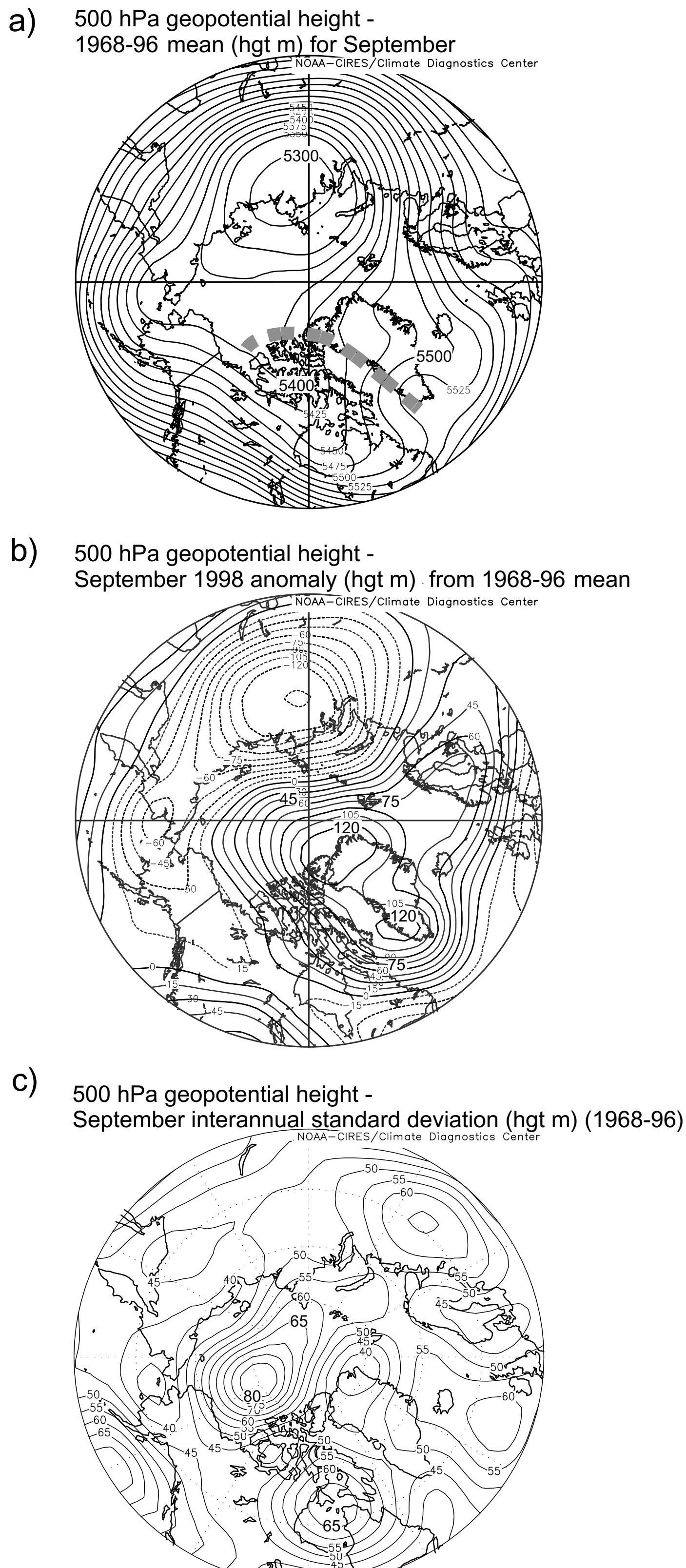

Fig. 10 The extent of the 500 hPa pressure surface anomaly in September 1998 (units for plots are geopotential height in metres) using the NCEP/NCAR reanalysis monthly data. a) Mean geopotential height values for that month show the location of the ridge (indicated by the heavy dashed line) over the QEI extending to western Greenland and Baffin Bay. b) Anomaly values for September 1998 compared to the September 1968–96 climatology showing the magnitude of the anomaly for September 1998. c) Monthly standard deviation of the mean September geopotential heights, 1968–96. Anomalies may be considered unusual events if they exceed the standard deviation. Such anomalies are present over the QEI and Baffin Bay.

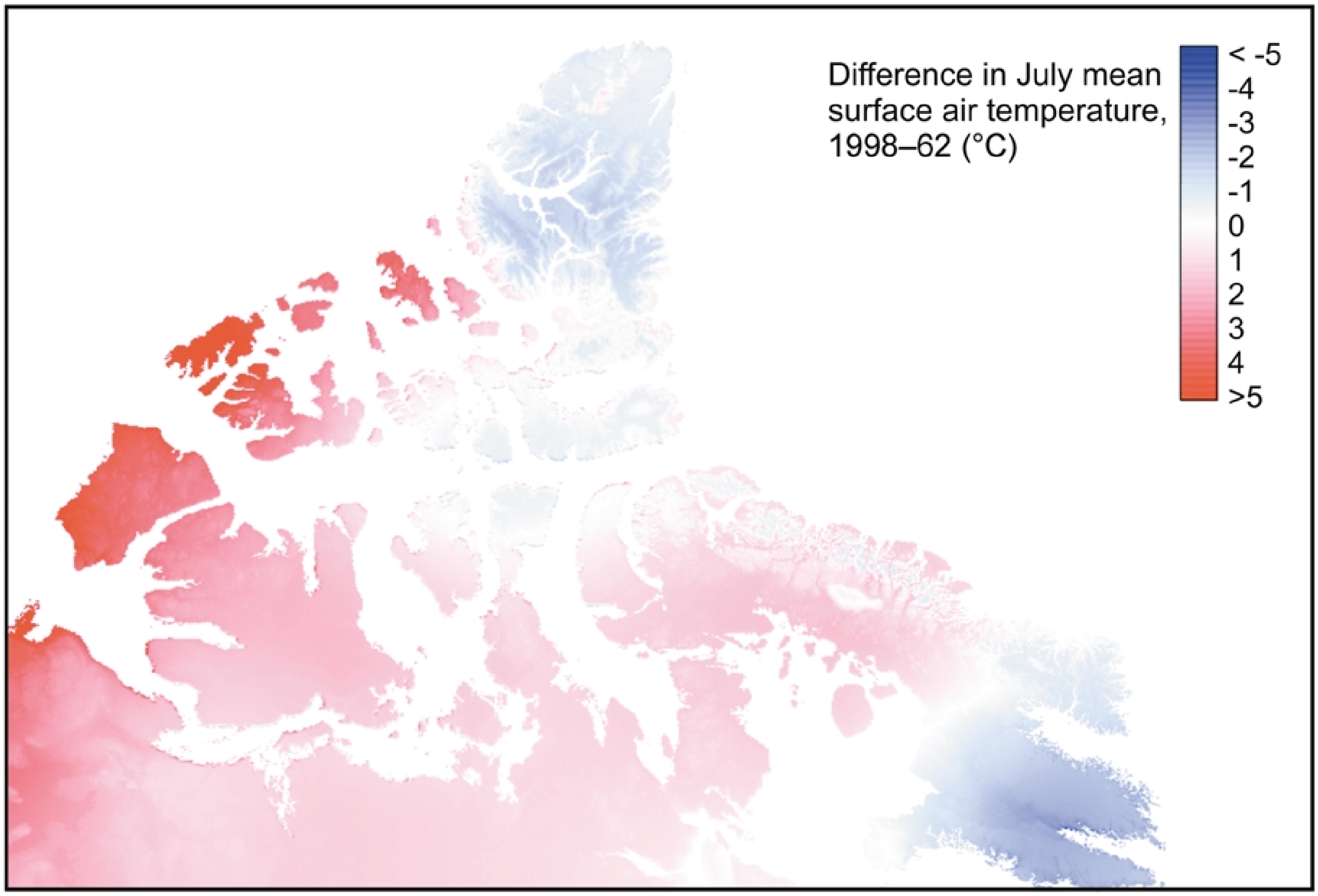

Fig. 11 July mean temperature (2 m) difference 1998 minus 1962, calculated using a detailed topoclimatic model at 1 km spatial resolution: red tones-1998 warmer than 1962; blue tones-1998 colder than 1962. Domain is the entire CAA. Units are degrees Celsius.

along the north-west edge of the QEI. A detailed analysis of their break-up sequence revealed that the removal of the plugs involved a combination of early ice melt south of the plugs generating open water, and the passage of two low pressure systems late in the melt season that provided the wind stress to fracture the ice (Fig. 16). This critical sequence of events (early melt, open water, high winds) was required to move the plugs, which explains why they disappear in some warm years, and not in others. An unexpected finding of the detailed study of the second storm (21 – 28 September) was that the persistent southerly winds in 1998 prevented the import of Arctic Ocean ice through the plug areas. These southerly winds also severely limited the amount of Arctic Ocean ice able to penetrate the CAA through other areas along the north-west edge of the QEI, where the landfast ice barriers also fractured during the summer of 1998. The Sverdrup plug was observed to break-up again in 1999 and 2000 (the Nansen plug remained intact), allowing the import of Arctic Ocean ice into the QEI. Time series of ice conditions in the western CAA (not shown) reveal that the concentration of multi-year ice remains low for 2–5 years after major clearing events such as occurred in 1962, 1970, 1981 and 1998. This is the time required for the infiltration of old ice into vacated areas from the zone of very thick multi-year ice found along the north-west coast of the Archipelago. Thus, heavier ice conditions within the Northwest Passage may occur several years after warmer Arctic summers such as that of 1998 (Jeffers et al., 2001; Agnew et al., 2001 and Melling, 2002). This cyclic pattern has important implications for Arctic shipping (Melling, 2002).

Investigation of the relationship of regional sea-ice conditions and sea-ice motion to atmospheric circulation indices such as the Arctic Oscillation (AO) and Pacific Decadal Oscillation (PDO) were inconclusive, in agreement with Maslanik et al. (1999) who found no clear correlations between Beaufort Sea ice severity and ENSO or the North Atlantic Oscillation (NAO). 1998 was not part of the much discussed 1990s positive AO and PDO period associated with recent high latitude warming over the northern hemisphere (Rigor et al., 2001; Proshutinsky and Johnson, 1997; Maslanik et al., 1996; Trenberth and Hoar, 1996; Phillips, 1995). However, the strong ENSO event in the winter of 1997–98 may have combined with a reversal in AO (and PDO) to produce the anomalous circulation over the Canadian Arctic in the summer of 1998, as indicated previously. Agnew and Silis (1995) discuss sea-ice response to atmospheric circulation at large scales (i.e., 500 hPa elevation) via storm-track and air mass variability.

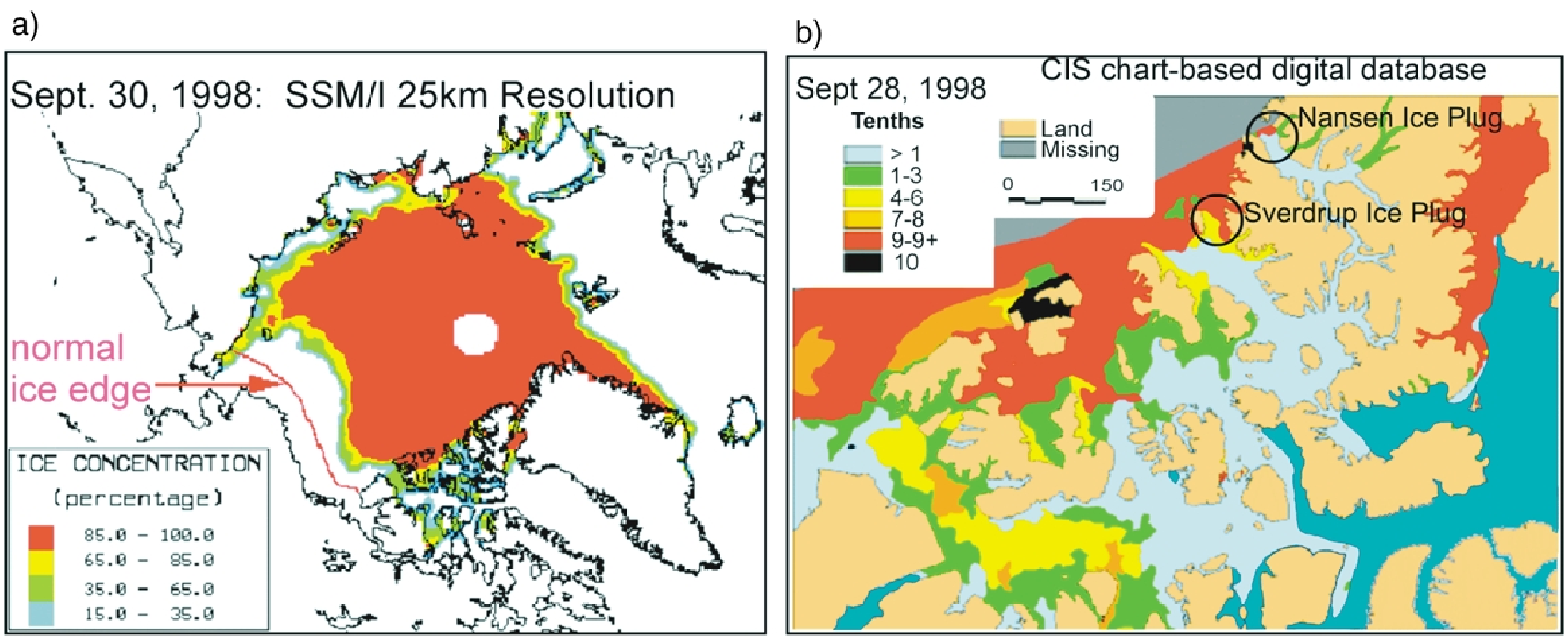

Fig. 12 a) SSM/I-derived sea-ice concentration for 30 September 1998 showing the normal ice edge position in the Beaufort Sea. b) Ice concentration in the QEI for 28 September 1998 from CIS digital ice charts. The dark blue area represents the 30-year median maximum open water (minimum ice) extent. The locations of the Nansen and Sverdrup ice plugs are indicated.

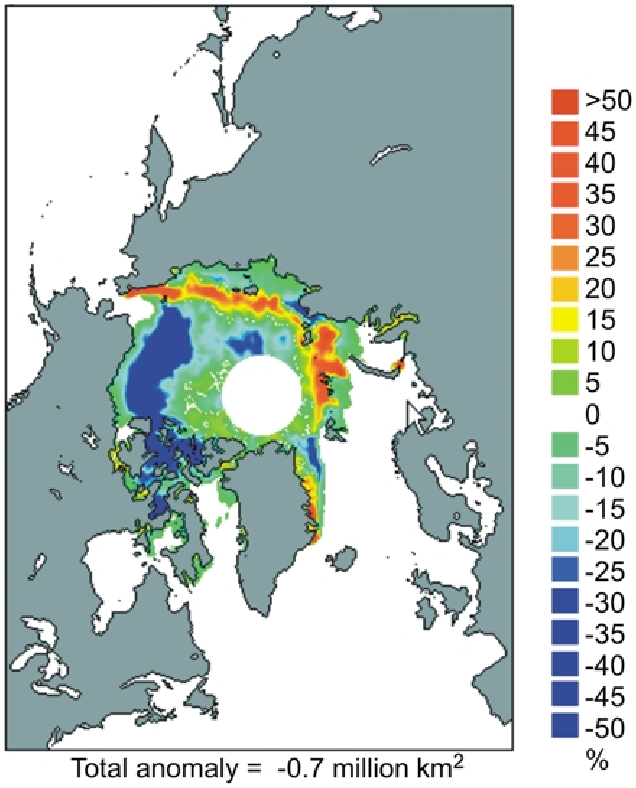

Fig. 13 Sea-ice concentration anomaly plot, September 1998. The anomaly concentration is shown as a percentage. Blue tones (negative values) indicate areas where September 1998 sea-ice concentrations were less than the September mean (Comiso, 1990).

b *Lake Ice*

The freeze-up and break-up of lake ice has been shown to have significant statistical relationships to surface air temperature at local to regional scales (Barry and Maslanik, 1993), and lake-ice cover can be readily monitored by satellite in both the visible and microwave portions of the spectrum. This provides the potential for monitoring changes in spring and fall temperatures over data sparse high latitudes. There was only one lake north of 60°N (Upper Dumbell Lake on northern Ellesmere Island) with directly observed data available for the summer of 1998 (thickness only). Time series of maximum ice thickness at this site (Fig. 17) revealed that 1998 had the second thinnest maximum ice thickness after 1990 during the 1960–98 period. The CIS initiated a program of regular satellite-based weekly monitoring of lake-ice cover across Canada in 1995. The program started with 34 lakes, increasing to 136 by 2002. Analysis of these data revealed that the shortest lake-ice cover seasons during the 1995-2002 period occurred in 1997–98 and 1998–99.

The warm summer conditions in 1998 affected ice-off dates (melt) and the subsequent freeze-up dates. Preliminary results from seven years of data suggest that lakes of different sizes within the same ecological zone have a different time response (lag) to climate conditions (Rouse et al., 2005). As an example, for Great Slave Lake early break-up occurred during the high sun season, which allowed the low albedo water to absorb more solar radiation during that year. It is this radiation that heats the lake, promotes large evaporation in fall and early winter and, at least in part, retards the autumn freeze-back rate. The early break-up and late freeze-up of Great Slave Lake in 1998 (Fig. 18) had the effect of prolonging the period of maximum sensible and latent heat fluxes, resulting in total evaporation that exceeded that of the other two (1997 and 1999) warmer than normal years by 25% (Rouse et al., 2003b).

SSM/I 85 GHz passive microwave satellite data have been used to map open water extent on Great Slave and Great Bear lakes since 1988 (Walker et al., 1997). The image resolution is relatively coarse (12.5 km), which means this approach can only be applied to large lakes. In 1998, break-up occurred 17 days earlier than average at Great Slave Lake, and 12 days earlier at Great Bear Lake (not shown). Freeze-up occurred 23 and 19 days later than average for each lake respectively. The 1998 freeze-up and break-up anomalies resulted in an extension of the open water season by 40 days. The composite long-term freeze-up series for Great Slave Lake, that includes historical in situ observations (Fig. 2e), indicates the latest freeze-up in the 40-year record probably occurred in 1998.

c *Snow Cover*

Snow cover has the largest spatial extent of any component of the cryosphere in Canada, and exerts a significant influence on climate and hydrology through modification of energy and moisture transfers and the storage of water. At local to regional scales, snow cover exhibits large spatial variability related to a host of factors such as elevation, exposure, topography, vegetation, and proximity to water bodies and storm tracks. These local-to-regional scale differences in the amount and timing of snow accumulation can contribute to differing cryospheric responses.

The summer of 1998 was characterized by a rapid, early retreat of the snow line across North America in April, May (Fig. 19) and June, with below-average snow cover over almost the entire area north of 60°N from June to September. This was, in part, due to a shallower initial snowpack (not shown), which is consistent with an Arctic-wide trend in evidence since the mid-1970s (Fig. 2b), as well as to the warmer temperatures. At Resolute Bay, snow cover disappeared three weeks earlier than normal (1955–99). The early disappearance of snow cover likely amplified the initial warming over land areas by providing more time for ground thaw, ground ice melt, and evaporation. The western Canadian Arctic experienced the warming earliest, resulting in the earliest recorded spring snowmelt during the 1955–99 period (Fig. 20). The eastern Canadian Arctic experienced warming into the fall, which was associated with the latest start to the snow cover season during the 1955–99 period. The early disappearance of snow cover in the Arctic in 1998 is part of a hemispheric trend (Brown, 2000) that also occurs in other components of the cryosphere, such as freshwater ice (Magnuson et al., 2000). This phenomenon is consistent with the observations of

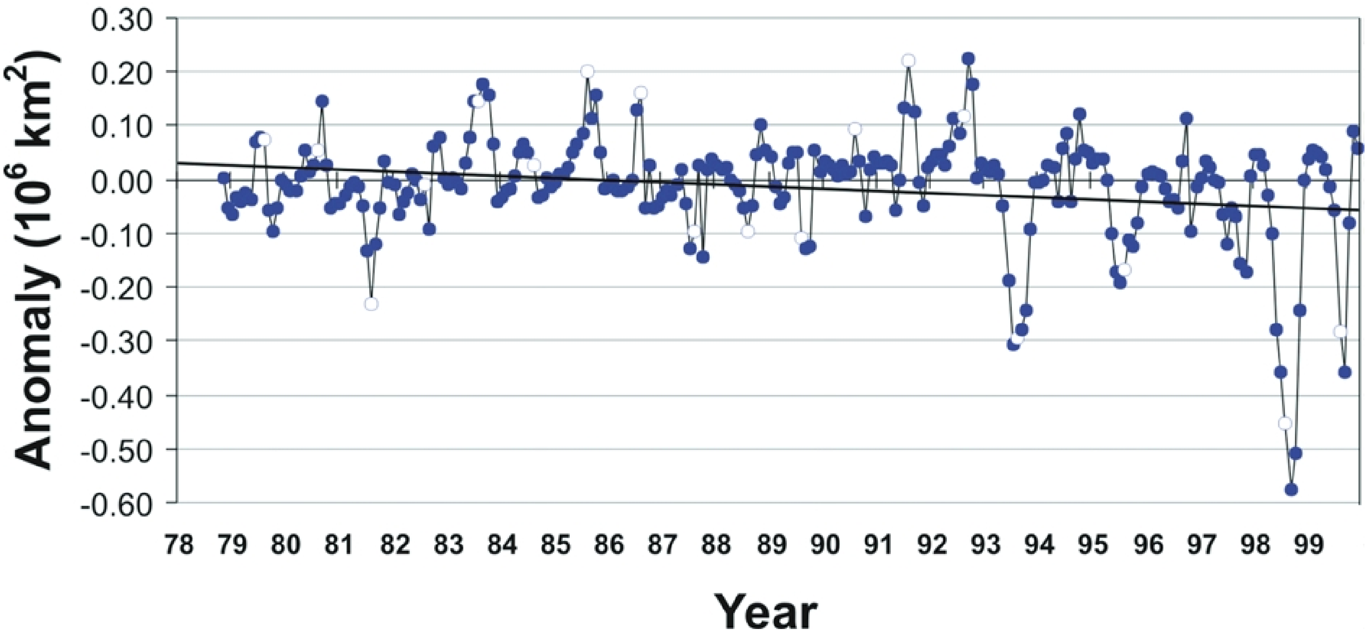

Fig. 14 Sea-ice area anomaly (deviation from the 1961–90 period) for the western Arctic Ocean (Chukchi Sea, Beaufort Sea and CAA) from passive microwave data. Open circles represent the August data points and are meant to improve readability.

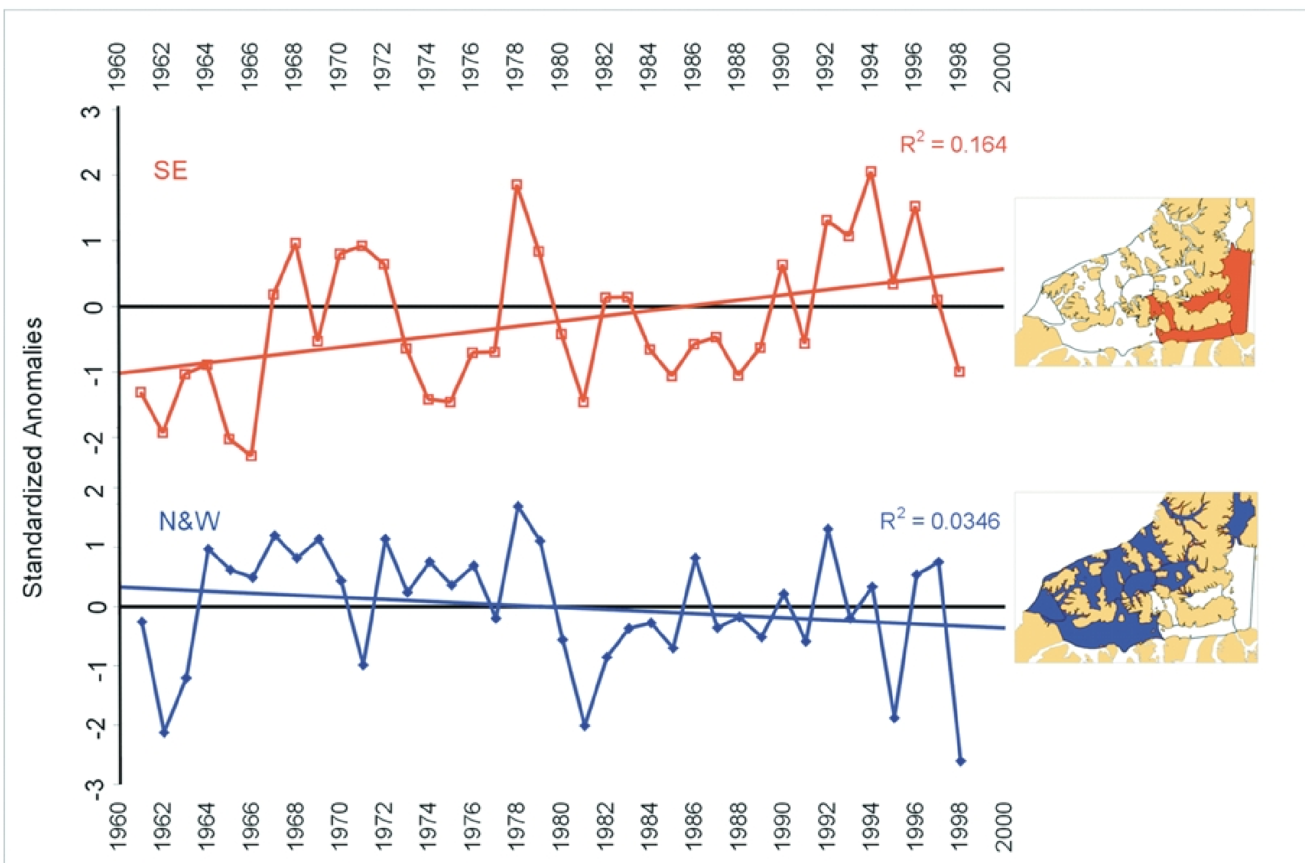

Fig. 15 Regional time series of sea-ice area at the height of the open water season, between the south-east zone (inset, top) and north and west zones (inset, bottom) of the QEI. The decrease in sea-ice area in the south-east is statistically significant at the 95% level.

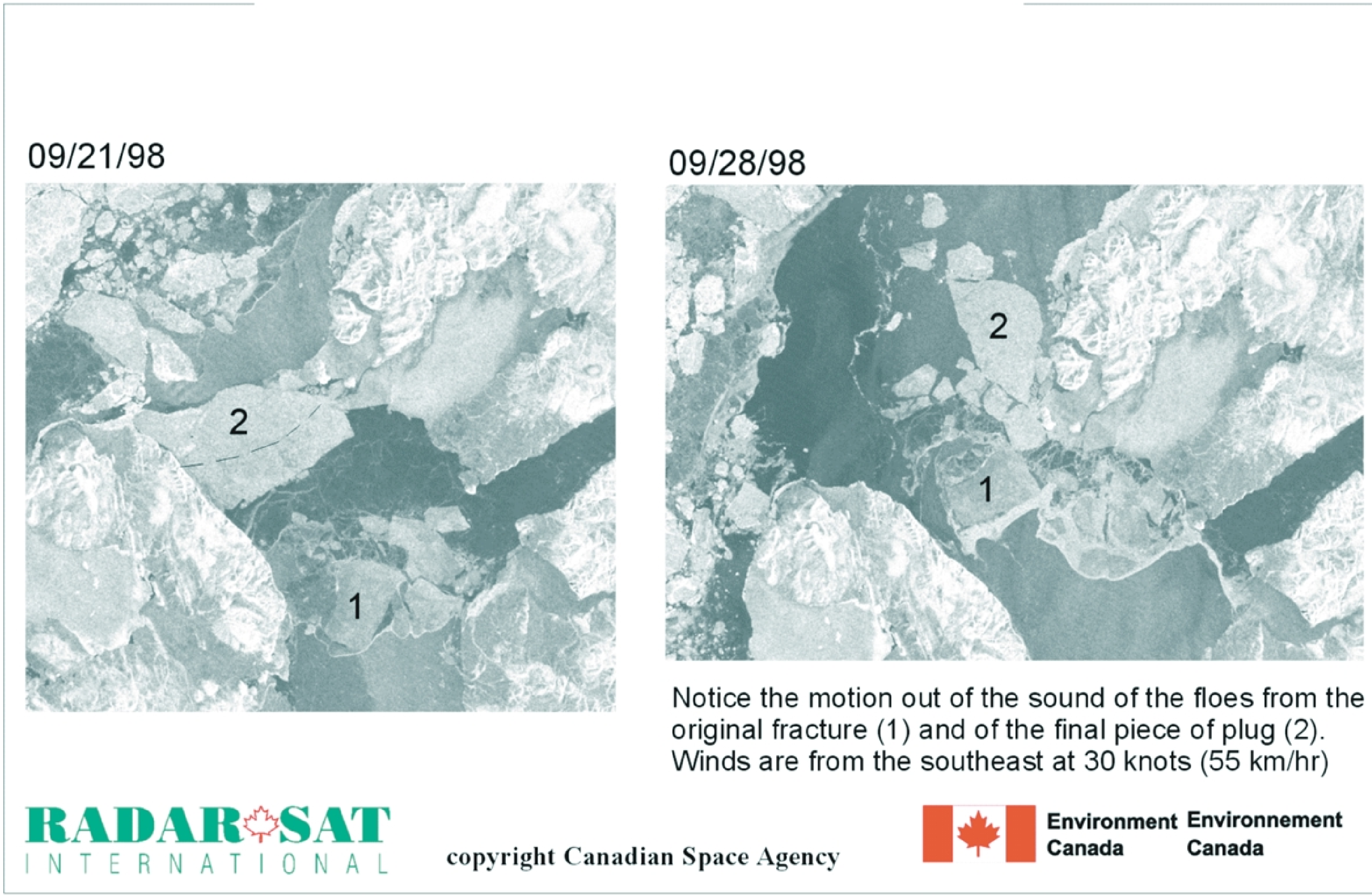

Fig. 16 Break-up of the final piece of the Nansen ice plug (northern Nansen Sound) during a storm that passed over the area on 23 September 1998 accompanied by strong south-easterly winds. Notice that all the pieces of the plug move north into the Arctic Ocean.

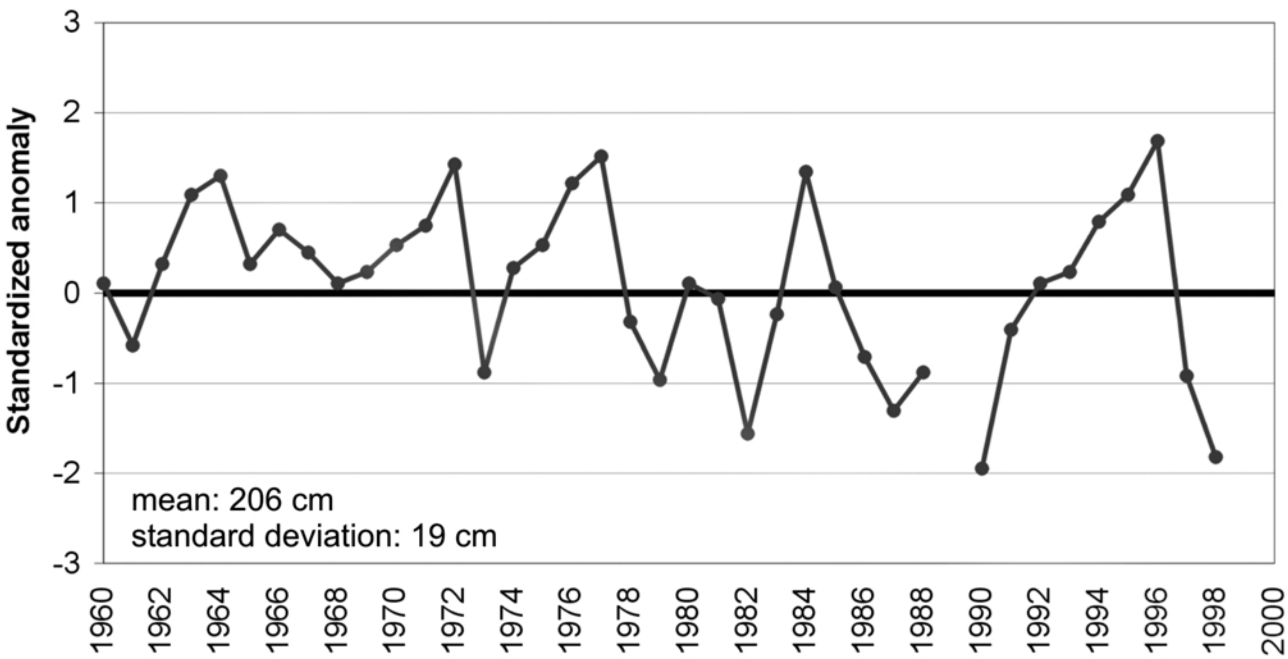

Fig. 17 Annual variation in maximum ice thickness at Upper Dumbell Lake near Alert, Ellesmere Island. Values are normalized with respect to a 1968–98 period. Overall depth mean and standard deviation (cm) are provided.

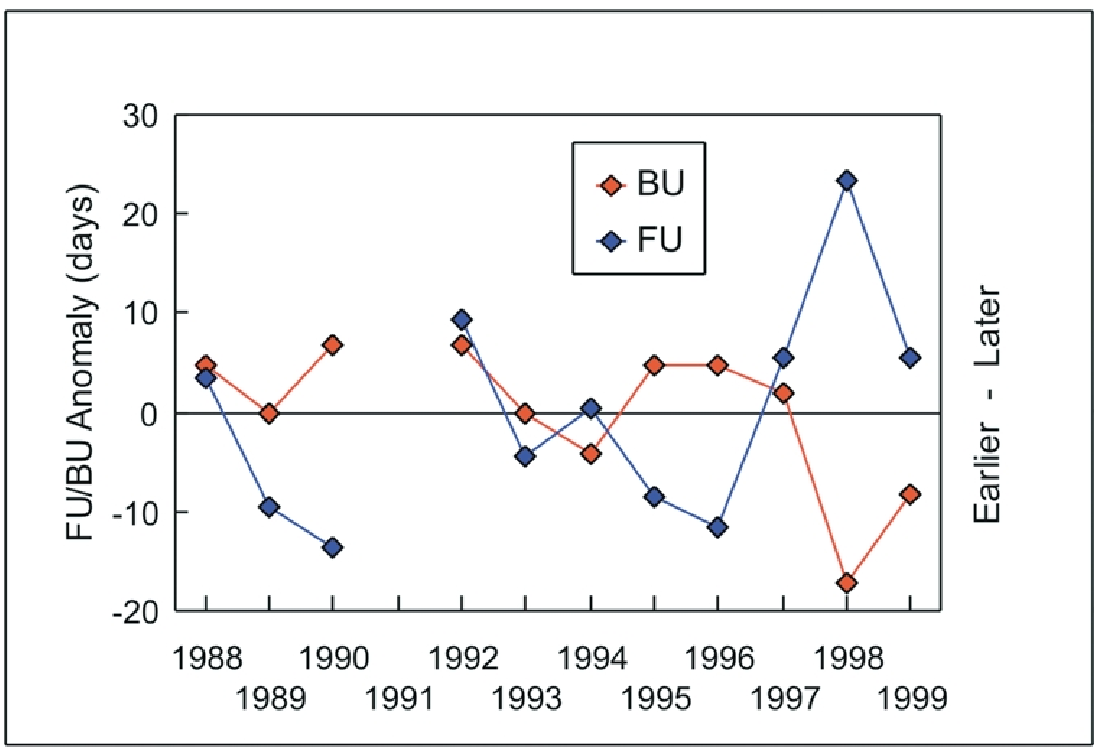

Fig. 18 Annual anomaly in freeze-up (FU) and break-up (BU) dates at Great Slave Lake derived from SSM/I satellite data. Anomalies are computed with respect to the mean for the entire period of data.

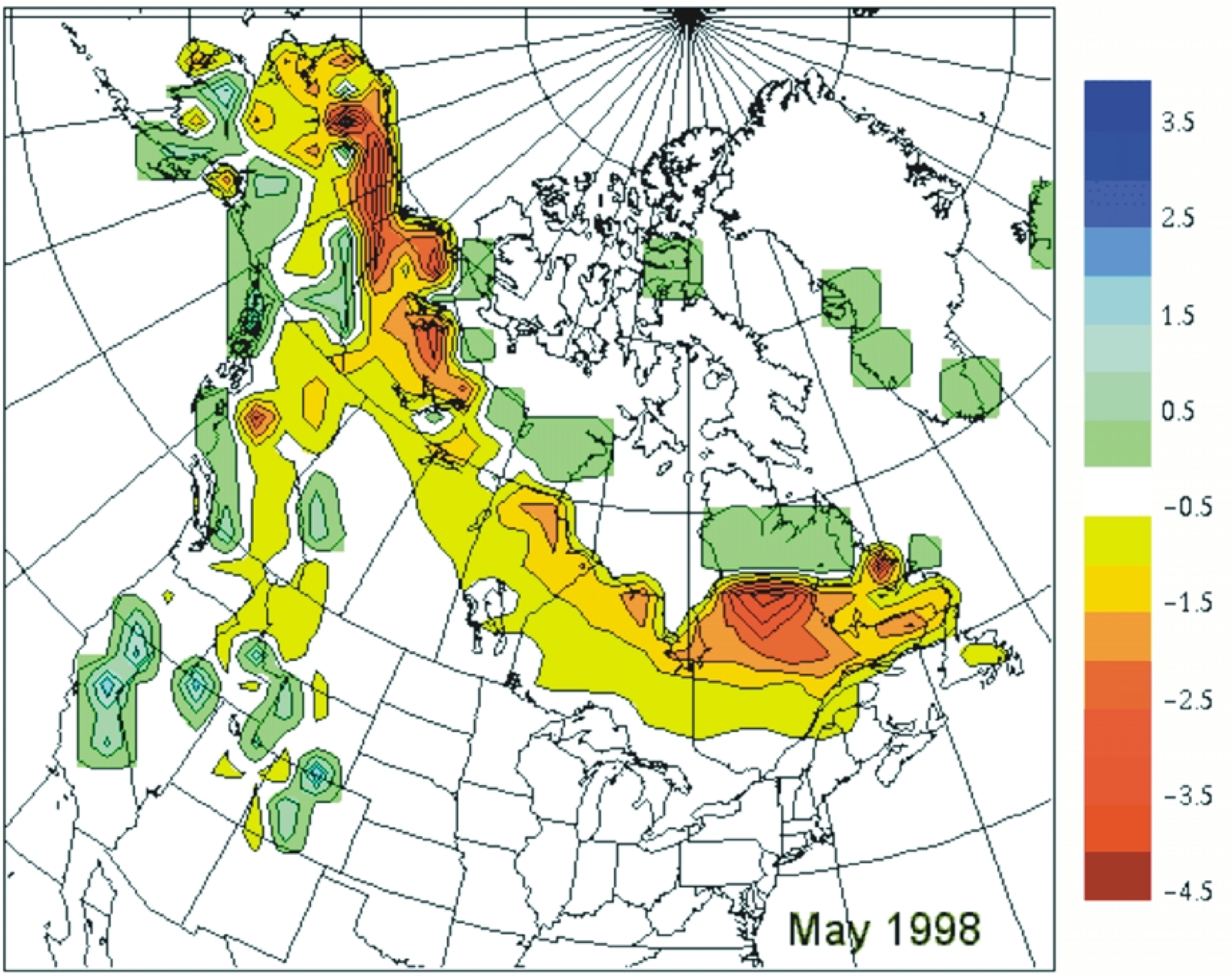

Fig. 19 Snow cover extent anomaly for May 1998 derived from the National Oceanic and Atmospheric Administration (NOAA) weekly satellite-derived product. Normalized anomalies are computed with respect to a 1972–95 reference period. Values greater than ±2 represent statistically significant departures from normal snow cover.

Groisman et al. (1994) that snow cover provides the strongest feedback to the Earth's radiation budget in the spring.

The hydrological implications of an extremely early snowmelt year over a tundra environment were studied as part of this project. Results indicated the importance of high rates of melt from semi-permanent snow banks (Edlund et al., 1989; Young and Lewkowicz, 1990) in maintaining locally higher than expected water tables in an otherwise dry tundra environment (Young and Woo, 2000, 2003). Snow was also observed to have a complementary effect on the ground temperature response in 1998. Heavier than normal snow accumulations during 1997–98 in some areas such as Alert, Ellesmere Island, acted to insulate the soil and contribute to warmer ground temperatures (Smith et al., 2001a, 2003).

d *Glaciers and Ice Caps*

The permanently ice-covered area of the CAA represents more than 60% of the ice-covered area in the Arctic outside Greenland. 1998 was characterized by a highly negative mass balance on the Melville Ice Cap (the most negative in the 1963–99 period) and the Devon Ice Cap (second most negative after 1962 in the 1961–99 record). Drambuie and John Evans glaciers, east of the Ellesmere Island Mountains, experienced high melt but not the highest in their short records (the Drambuie Glacier record began in 1977; the John Evans Glacier record began in 1996). In contrast, the Meighen Ice Cap (record starting in 1960) had a slightly negative mass balance (above normal melt) in 1998 and similar near-normal mass balances were observed on White and Baby glaciers on Axel Heiberg Island. The variation in mass balance across the QEI reflects the strength of the temperature anomaly gradient towards the west (Fig. 3, 1998), as seen on the Melville Ice Cap (Figs 21 and 22), and the importance of individual summer storms, as demonstrated by the low-pressure system which tracked into the CAA from the Arctic Ocean (Alt, 2001). This event resulted in early snowfall (late

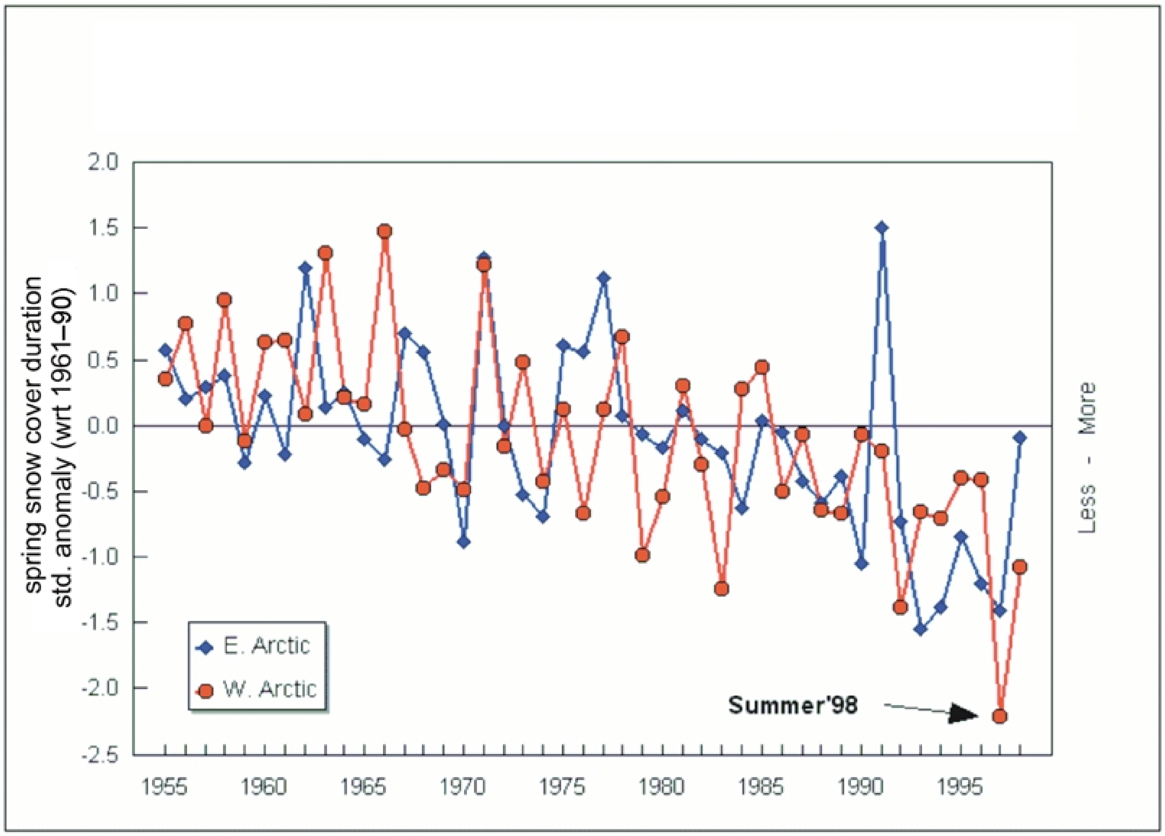

Fig. 20 Annual variation in spring snow cover duration over the eastern and western Canadian Arctic (either side of 100°W) derived from in situ daily snow depth observations.

July), ending the melt season on the Meighen Ice Cap (Fig. 22) and caused varying degrees of albedo increase on QEI glaciers, as illustrated by automatic weather station snow depth records from the Meighen and Melville ice caps and the John Evans Glacier (Fig. 22).

The mass balance (summer melt) time series show a stronger response to the warm summer of 1962 (Figs 2d and 21) than is seen in the sea-ice time series (Figs 2c and 15). Previous studies showed that high melt on QEI glaciers could be linked to a strong ridge over the CAA and a shift of the polar vortex to the Siberian side of the Arctic Ocean (Alt, 1987). All the ice caps that have records longer than 20 years (Devon, White, Baby, Drambuie, Meighen, Melville South) show overall negative balances: i.e., they are losing mass on a long-term basis. A trend to even greater negative balances, beginning in the mid-1980s, is becoming apparent in these records but is weakest in those ice bodies in the north-west (Meighen and Axel Heiberg).

Long-term paleoclimatic records have been developed from ice cores, using $\delta^{18}O$ for annual, and melt layers for proxy summer temperature indicators (Koerner, 1997). Various aerosols captured in the ice (e.g., pollen, dust, salt, major ions, trace metals) also provide information on past climatic conditions, as well as prevailing atmospheric circulation patterns if the source regions are known (Bourgeois et al., 2000, 2001; Fisher and Koerner, 2003; Gajewski and Atkinson, 2003). Such records reveal that Canadian Arctic ice caps were at their maximum extent during the last 10 000 years, during a recent cold period which lasted from approximately 1400 to 1800 CE in the QEI (Gajewski and Atkinson, 2003). In contrast, temperatures were substantially warmer than present in the High Arctic between about 10 000 and 6 000 years ago. The paleoclimatic record also provides evidence that past warming events had a regional pattern similar to recent trends, with stronger warming over the western Arctic (Koerner and Alt, 2001).

Preliminary analysis of a 20-year pollen deposition record from the Agassiz Ice Cap showed a maximum peak in local (for the most part herbaceous) pollen in the late summer–early fall layer of 1999. This may have been due to the long, warm summer of 1998 which allowed a greater than normal number of flowers to set.

e *Permafrost*

Permafrost (perennially frozen ground) underlies almost half of the Canadian landmass, much of it in areas that are expected to experience considerable warming in the future (Smith and Burgess, 1999; Smith et al., 2001b). The response of the ground thermal regime to atmospheric warming depends on local variations in vegetation, surficial materials, moisture content, wind exposure, and snow cover. For example, warm winter ground temperatures at Alert during 1997–98 were at

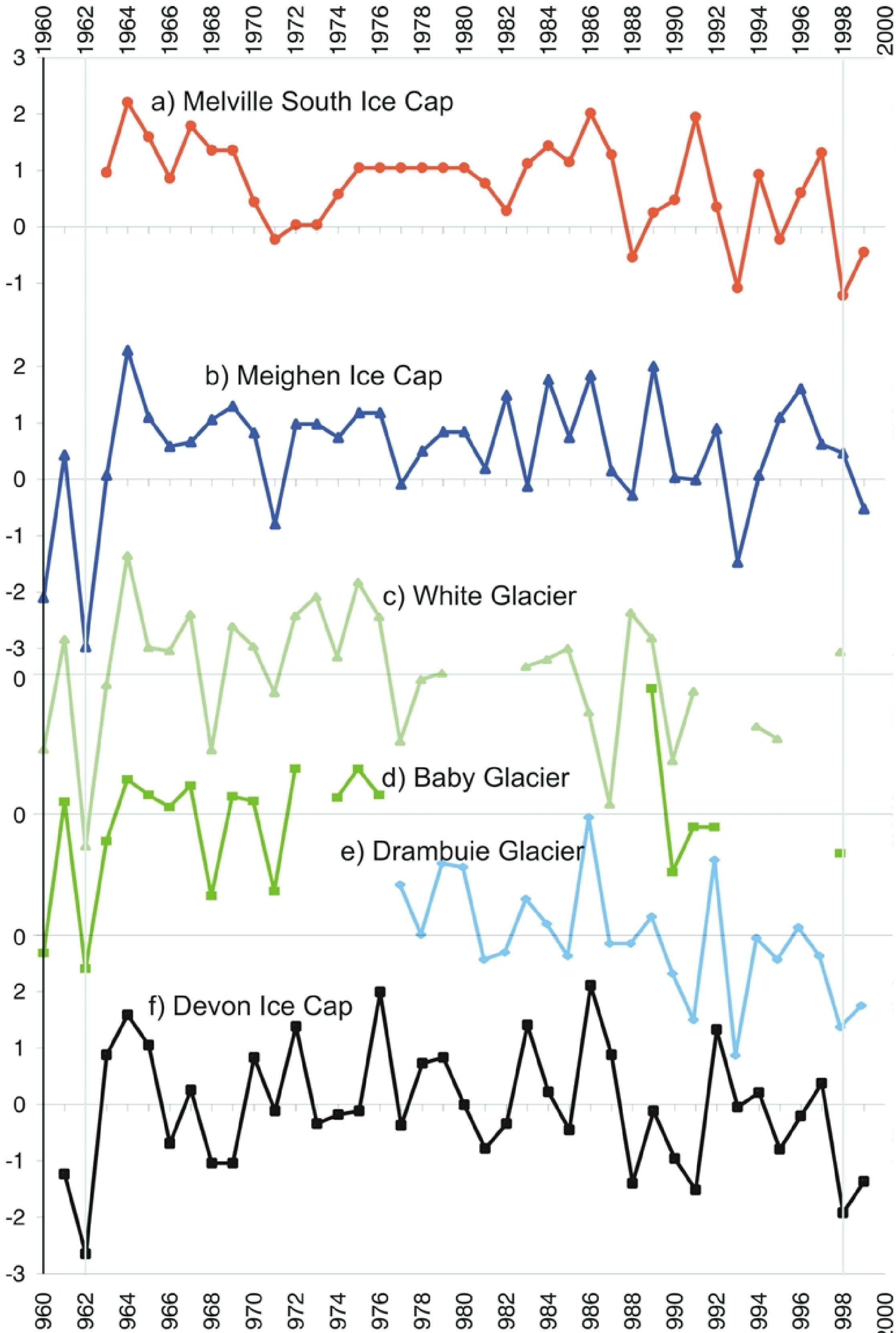

Fig. 21 Annual mass balance: a) Melville South Ice Cap (data missing 1974–80; average value used), b) Meighen Ice Cap, c) White Glacier, d) Baby Glacier, e) Drambuie Glacier, and f) Devon Ice Cap. Values are standardized anomalies calculated relative to 1968–98 wherever possible. 1962 and 1998 are marked for comparative reference (Cogley et al. 1995).

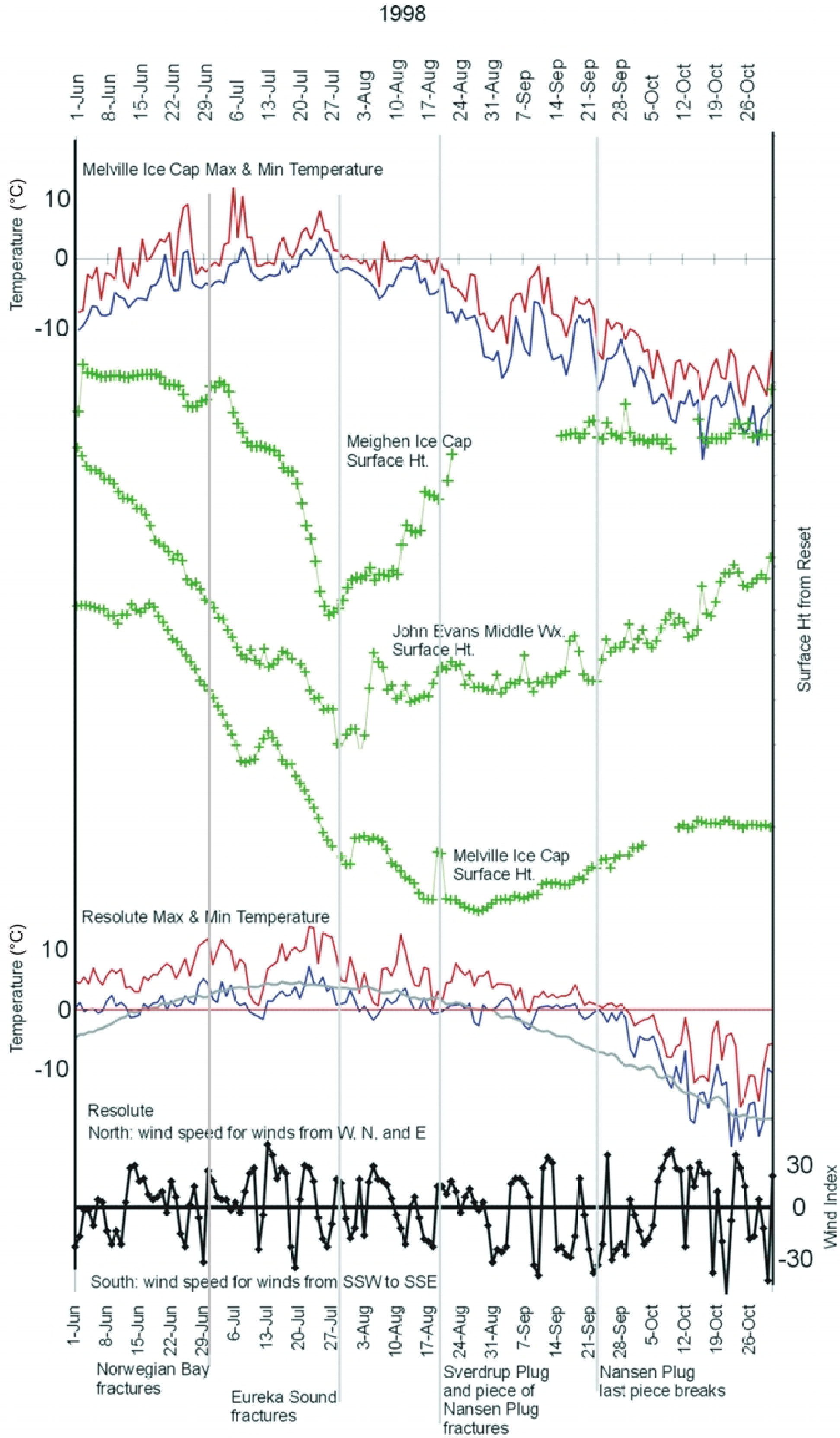

Fig. 22 Daily maximum and minimum temperature (°C) at Resolute Bay and Melville Ice Cap, wind index from Resolute, and surface elevation on Meighen, Melville and John Evans glaciers relative to the height at the time of reset for 1998. The "wind index" represents wind speed as positive if it is from the north quadrant and negative if it is from the south quadrant. Units are km h^{-1}. Some major sea-ice events are indicated.

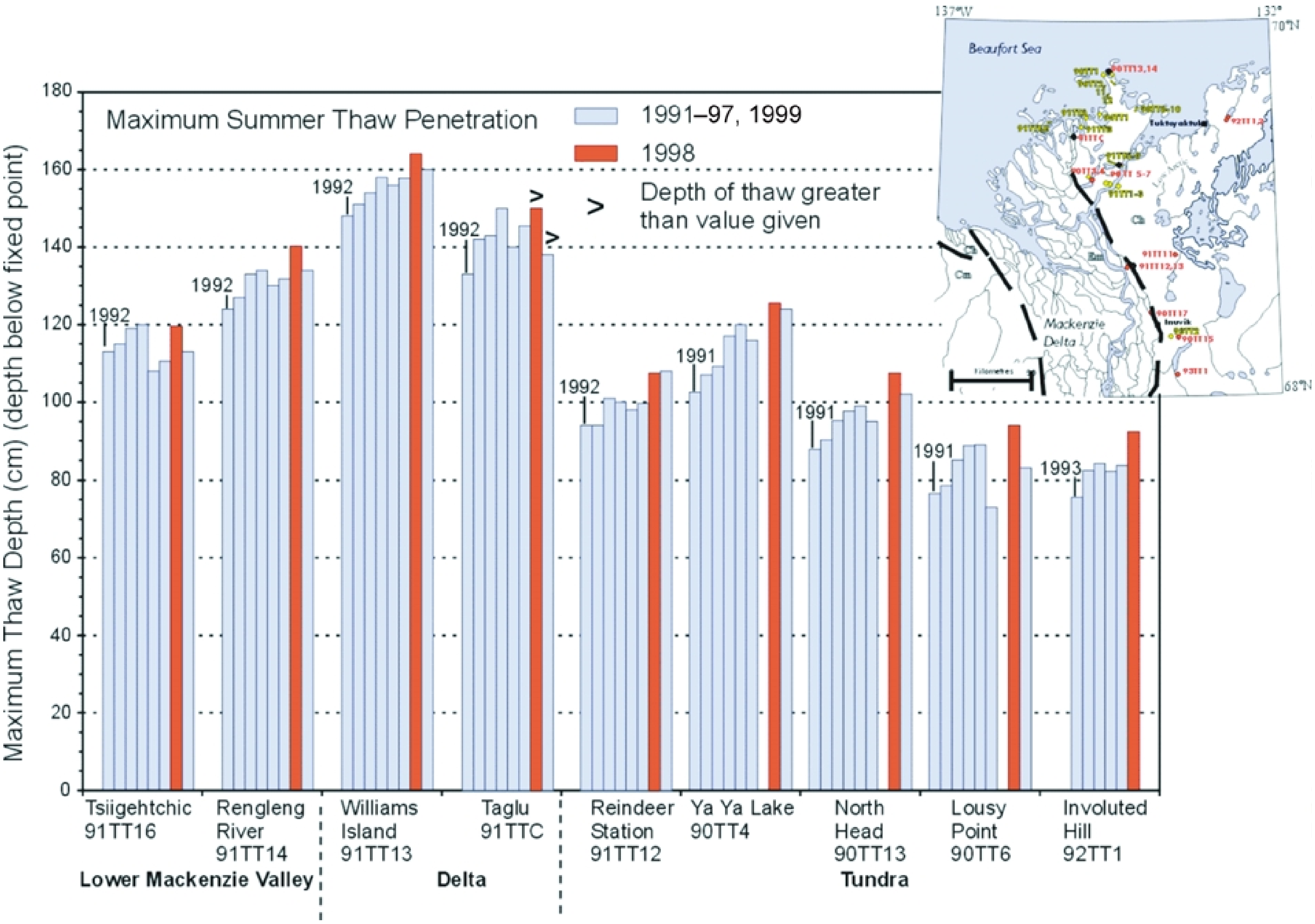

Fig. 23 Maximum summer thaw penetration determined for nine sites in the Mackenzie Delta and Tuktoyaktuk Peninsula region between 1991 and 1999 (reproduced from Smith et al. (2001a) with permission of Natural Resources Canada).

least partially due to an anomalous deep winter snow cover (60 cm maximum compared to a normal of ~20 cm (Smith et al., 2001a, 2003; Romanovsky et al., 2002)).

Wolfe et al. (2000) and Smith et al. (2001a) reported on the state of the active layer in 1998 based on thaw tube observations and ground surface temperature measurements made since 1991 across the northern Mackenzie Delta (Fig. 23). The greatest thaw depth (measured relative to a fixed point above the ground surface) on record occurred in 1998, with thaw depths typically 10 cm greater and ground subsidence 1 to 7 cm greater than the previously recorded maxima. Thicker active layers were generally observed in 1998 (Brown et al., 2000). At sites with ice-rich soil, however, the large increase in thaw penetration was accompanied by significant ground subsidence and hence very little change in active layer thickness. The thaw season ranged from 8 to 25 days longer than in previous years, with ground surface temperatures 1° to 4°C higher. Ice wedge observations indicated recent deeper thaw penetration, but this may be a cumulative effect and not solely due to the conditions in 1998 (Wolfe et al., 2000). Shallow ground temperatures have been measured at Baker Lake since 1997, and thaw depth increased during the monitoring period with the largest increase occurring in 1998. Warming occurred later at Baker Lake compared to the western Arctic, with warmer ground temperatures in the late summer and early fall resulting in an extension of the thaw season in 1998 (Smith et al., 2001a; Brown et al., 2000).

The magnitude of the ground ice warming in the CAA during 1998 is evident in the long-term July 100 cm ground temperature record at Resolute Bay (Fig. 2f), showing that the warmest ground temperatures since 1965 likely occured in 1998. Field studies showed that these warm temperatures produced enhanced ground ice melt in the area. Documentation of active layer detachment slide (ALDS) activity at three sites on Fosheim Peninsula, Ellesmere Island (Lewkowicz, 1990, 1992; Harris and Lewkowicz, 2000), underway since 1988 (and estimated for earlier periods), suggests a possible increase in activity since 1975 compared to the previous 75 years. Indications are that in 1998, 10 new slides were initiated at Big Slide Creek (Fig. 24), the most since records began.

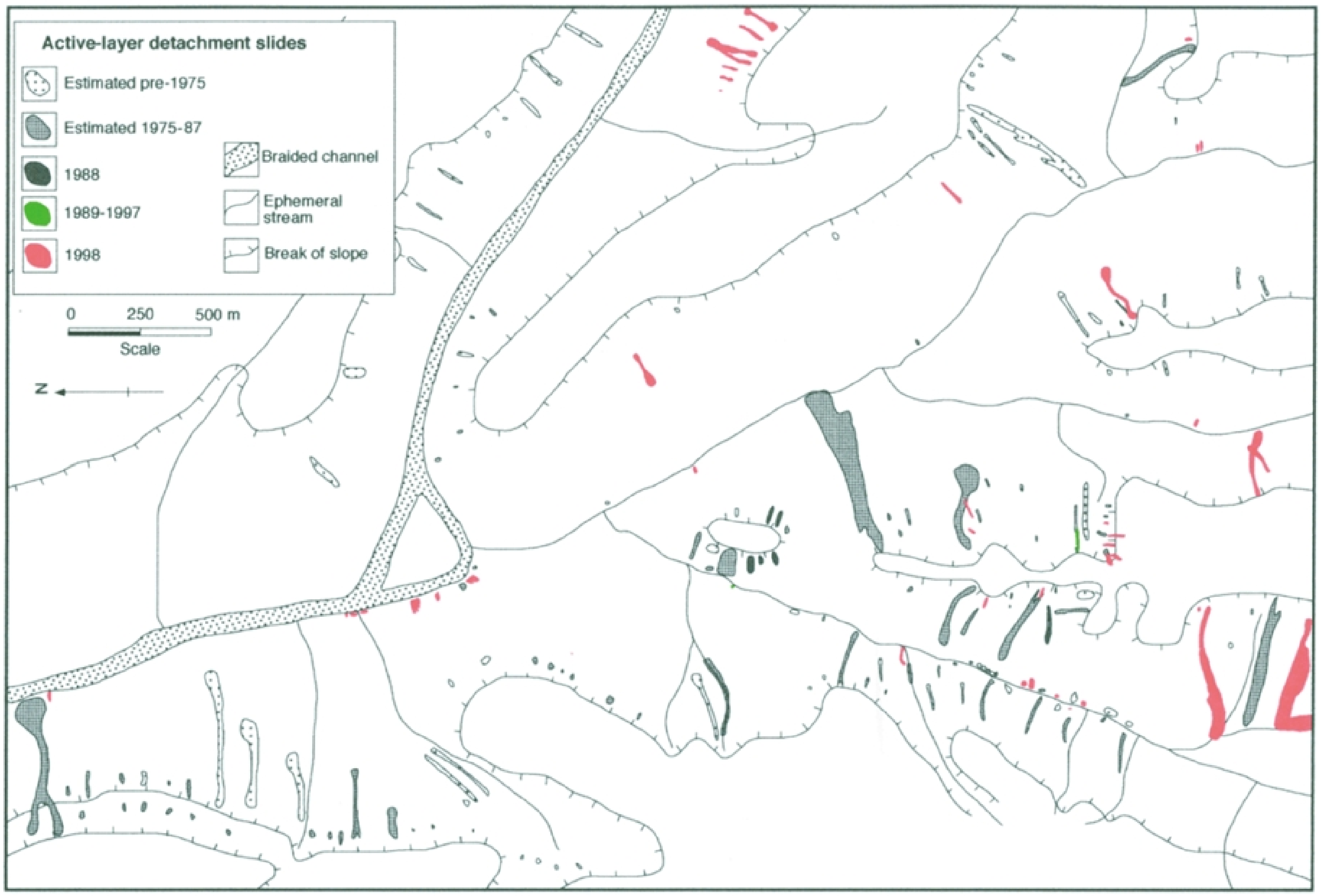

Fig. 24 Map of active layer detachment slides in "Big Slide Creek" on Fosheim Peninsula, Ellesmere Island. Red slides are those from 1998 (green from 1989–97, dark grey 1988, light grey estimated 1975–87, and dotted estimated pre-1975).

However, lower activity at the other two sites made 1998 the second most active year (after 1988) during the 1987–2000 period for the three sites combined. It was hypothesized (Harris and Lewkowicz, 2000) that an extreme precipitation event late in the preceding summer contributed to the 1998 active layer detachments, by promoting the development of ice lensing near the base of the active layer.

4 Factors affecting cryospheric response

Attempting to piece together an integrated cryospheric response to the extreme warming event is complicated by the differing spatial and temporal scales of the processes and feedbacks. Nevertheless, this study was able to document a number of factors that were significant in determining the response of the cryospheric elements to warming. These can be generally classified as *preconditioning, critical events*, and the role of *local-scale differences in the physical environment* (e.g., topography, soil properties, snow accumulation, vegetation cover, etc).

a *Preconditioning*

Several examples of the importance of preconditioning on cryospheric response to anomalous warming were noted in this project. In the Beaufort Sea, anomalous south-easterly winds cleared sea ice far to the north-west by early May, which permitted unusual warming of the upper ocean with a positive feedback to air temperature and to sea-ice melt. In the Mackenzie Delta, the warmth of the winter of 1997–98 and preceding summer were likely contributors to the extensive ground thaw that occurred during the spring of 1998. In the spring of 1999, lake ice broke up early due to late freeze-up in the fall to early winter of 1998. Similarly, in the CAA, the extensive open water at the end of the summer of 1998 was followed by early break-up in both 1999 and 2000. Preconditioning of the active layer by a rainfall event towards the end of the previous melt season may have been partly responsible for the frequency and distribution of ALDS on northern Ellesmere Island during the warm summer of 1998.

b *Critical Events*

Individual weather events can have a large effect on the High Arctic cryosphere. For example, a strong southerly wind associated with a low pressure system was responsible for breaking the last of the Nansen Sound ice plug and for dislodging the Sverdrup Channel ice plug, and an early snowfall ended the melt season on some of the glaciers and ice caps in the QEI in 1998.

c *Local-scale Difference in the Physical Environment*

The response of the cryosphere to the warmth of 1998 varied not only regionally, but also locally due to differences in surface materials and local topography. For example, the effect of the longer melt season on total melt appears to have been more pronounced at the upper station on the John Evans Glacier than at the middle and lower stations, probably because the location is favourable to chinook winds (Boon et al., 2003); considerable differences in thaw penetration, active layer thickness and ground thermal regime were observed depending on ground ice content, snow cover, and surface vegetation; and significant differences were observed in ground ice meltwater depending on soil type and proximity to melting semi-permanent snowbanks. Consideration of local site factors also helps to explain different temporal and spatial responses of ALDS.

5 1998 as a harbinger of things to come?

The occurrence of the extreme warming event in 1998, during a decade of likely unprecedented global warmth (IPCC, 2001), raises the question as to whether this is a harbinger of things to come if global warming is occurring. To place these results in the context of long-term global warming, however, is problematic because we are looking at the cryospheric response during a single extreme warm summer, as opposed to a future warmer mean climate with different boundary conditions. It is also difficult to extrapolate from a single year when cryosphere-climate processes operate over a wide range of different timescales (e.g., Bamzai, 2003). For example, cryospheric components, such as snow cover, respond very rapidly to a warming anomaly, while components such as ocean circulation, permafrost temperature and glacier mass balance respond with time lags ranging from years to decades. Factoring in the influence of modes of atmospheric variability, such as those expressed by the Arctic Oscillation, exacerbates the problems associated with attempts at extrapolation, given their non-linear influence on cryospheric response coupled with the unknown response of these modes to warming (Ostermeier and Wallace, 2003). The consideration of multiple scales of activity is also important in this regard, as they are with general considerations of a basic understanding of sea-ice dynamics (McNutt and Overland, 2003). Nonetheless, as was shown in this study, examining a single extreme year can yield important insights into cryospheric processes, and the results can be used to evaluate climate model simulations (e.g., spatial patterns, sequence of events, atmospheric circulation and intensity of anomalous conditions). Furthermore, Polyakov et al. (2002) stated the hypothesized occurrence of a polar amplification of global warming has not materialized, and that signal detection is made more difficult by the large degree of variability exhibited in the Arctic. A more complete understanding of extreme years will also help isolate fundamental trends from the noise of variability.

6 Conclusions and recommendations

The summer of 1998 was characterized by a significant response of most cryospheric variables (snow-cover area and duration, sea-ice concentration and extent, lake-ice thickness and duration, glacier melt, active layer thaw depth, permafrost temperature, ALDS occurrence and ice wedge melt), to the warm temperatures that covered a wide area stretching from the Mackenzie Delta to the CAA (Tables 1 and 2). The most distinctive characteristic of this summer was the extension of above-normal (and above freezing) temperatures well into the fall. This extended warming was largely due to the displacement of the polar vortex to the Siberian sector of the Arctic Ocean, which favoured the penetration of warm southerly air into the Canadian Arctic. The formation of new sea ice in the Beaufort Sea was delayed by three weeks, to early November.

While the warmest year of the 1950–2003 period in the Canadian Arctic was 1998, this project showed that similar responses were generated in some areas and in some components of the cryosphere in other years. The explanation for this follows logically from the project findings that preconditioning, critical events and local-scale differences in the physical environment are all involved in determining the cryospheric response to warming in the Canadian Arctic. These findings underscore the complexity of the Arctic cryospheric system, and demonstrate that warmer temperatures per se do not necessarily generate a predictable response because other processes are involved, such as the effect of snow cover on ground temperatures. In addition, model sensitivity studies of processes such as sea-ice freeze-up and break-up suggest that there is a strong time-dependent sensitivity to temperature forcing (Fig. 25). In this figure a fixed, anomalous temperature perturbation is applied to a one-dimensional sea-ice model at progressively later days throughout the year and the resulting change in melt season length is plotted. It can be seen that, before or after a certain Julian day, the introduction of extra (or reduced) thermal energy does not have an effect. When there is an effect, it is not uniform but is time dependent. If applied at the right time, the response is strong. For example, when applied near Julian day 165 the positive thermal anomaly results in an increase in an open water season length of ~25 days, whereas when the same anomaly is applied on Julian day 125, there is no change in the open water season length (Flato and Brown, 1996). The results of this study also point to a need for further process research to understand better the response of the cryosphere to warming, such as the role of ice lenses in ALDS, documenting the frequency of local chinook winds, and investigating the complex interactions between synoptic events and Arctic Ocean ice import. For example, Proshutinsky et al. (2002) identify patterns in the Beaufort Gyre interannual variability and link these to broader regional patterns of climate. Modification of the gyre could enhance, supress, or alter effects that have been described here.

One of the main contributions of this study was the development of time series of cryospheric parameters using satellite, historical, and proxy sources of data to place 1998 in the context of recent climate variability. This information, available to the public at the "State of the Canadian Cryosphere:

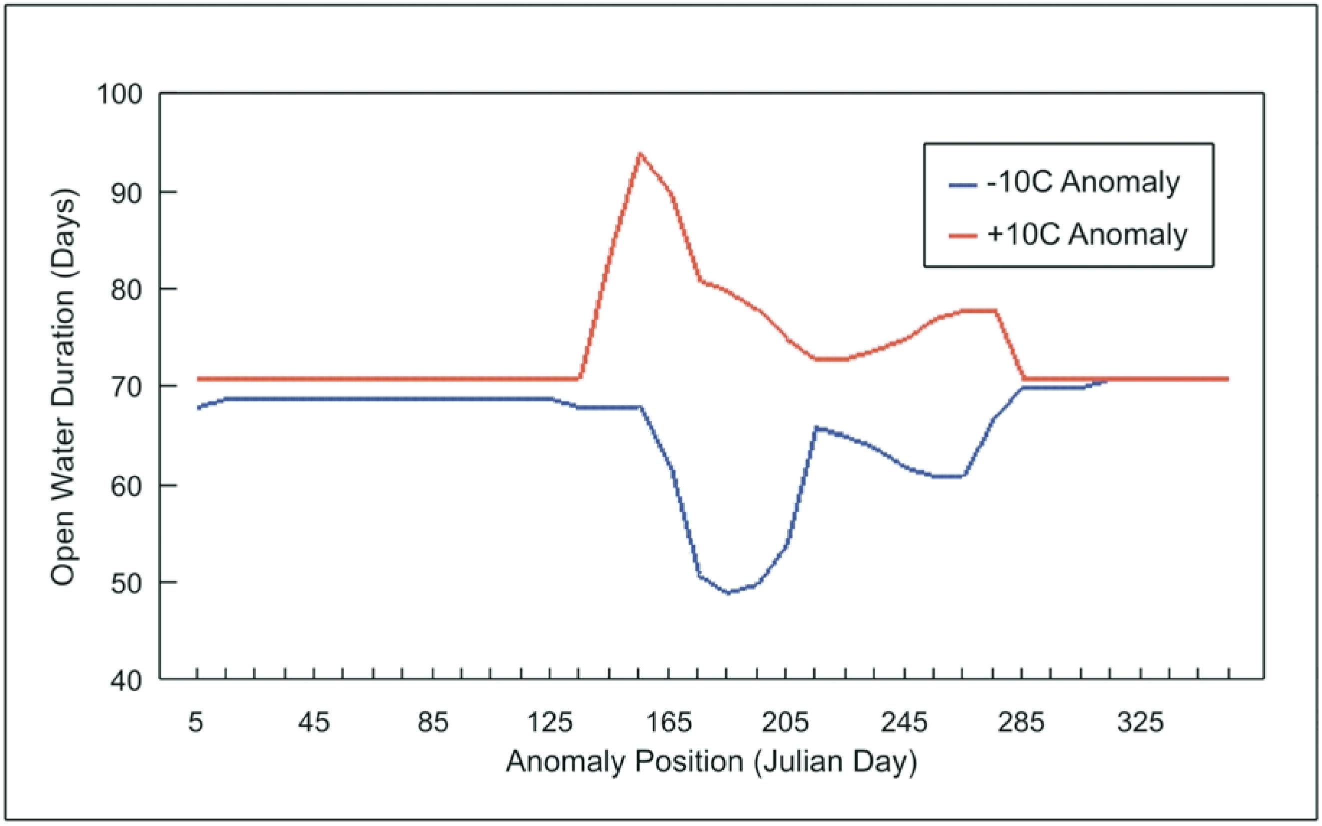

Fig. 25 Simulated response of open water duration to a moving 10-day 10°C air temperature anomaly based on the Flato 1-D thermodynamic sea-ice model used in Flato and Brown (1996).

Canadian Cryospheric Information Network" website http://www.socc.ca/examples/ccin/metadb/metaset.jsp, is vital for climate monitoring in the Arctic, and ongoing effort should be devoted to update, extend and assess the homogeneity of important Arctic cryospheric time series. The reduction of in situ observing networks in the Arctic during the 1990s was a major obstacle to documenting cryospheric variability, particularly for lake freeze-up and break-up, daily snow depth, and weekly fast ice thickness. Some of the observing network deficiencies were addressed in the Canadian government's "Action Plan 2000" initiative (snow depth, weekly fast ice, permafrost, glaciers), but there are still important thematic and regional gaps that need to be filled (Brown and O'Neill, 2002).

An effort also needs to be made to improve access to Arctic datasets, which are distributed over a number of organizations and locations. The CCAF-supported Arctic data rescue project, and the Canadian Cryospheric Information Network (CCIN) initiative at the University of Waterloo will help improve data access and availability. However, additional resources are needed to document, quality control, and update important Arctic datasets, and to make these available to the research community. The co-operation and involvement of the participants in this project was greatly facilitated by the Cryospheric System in Canada (CRYSYS) project (www.crysys.ca), which has developed an active network of Canadian cryospheric scientists since the early 1990s. The co-location of the 2001 CRYSYS meeting and the CCAF High Arctic Inventory and Rescue meeting provided a unique opportunity for a joint session on the summer of 1998 during which 22 presentations laid the groundwork for the final report. Sharing of the data and results from this project is being facilitated by the CCIN.

Acknowledgments

This project was made possible with funding support from the Climate Change Action Fund. The authors would also like to acknowledge the support and encouragement of Dr. Barry Goodison, CRYSYS Project Principal Investigator. The various investigations contributing to this study also received funding and in-kind contributions from the following sources: the CRYSYS project (Meteorological Service of Canada), the Department of Fisheries and Oceans including the Canadian Coast Guard, the Geological Survey of Canada and the Polar Continental Shelf Project (both divisions of Natural Resources Canada), the National Sciences and Engineering Research Council of Canada, the Federal Programme on Energy Research and Development. We appreciate review comments made by Bob Taylor (Geological Survey of Canada, Atlantic) and Kathryn Parlee (coordinator, Canadian, Climate Impacts and Adaptation Research Network, Coastal Zone), and three anonymous reviewers, one of whom made many useful content and style comments.

APPENDIX – Notes related to the derivation of Figure 2

The purpose of Fig. 2 was to provide a concise visual summary of Arctic temperature and cryospheric variability for a selection of variables with more-or-less complete data for the last four decades. Where possible, areal averages were used, but in several cases the individual data series were too fractured to permit this. A brief discussion of the derivation of each time series follows.

a *Air Temperature* (Fig. 2a)
This series was derived from the 2001 update of the Jones (Jones and Moberg, 2003) gridded land air temperature anomaly dataset (CRUTEM1v), adjusted to account for changing numbers of stations over time. Monthly averages were computed over the North American landmass north of 60ºN with cosine weighting applied to take account of decreasing grid-cell size with increasing latitude. The May to October average was computed from the monthly anomaly series and standardized with respect to the 1968–98 period.

b *Spring Snow Cover* (Fig. 2b)
This series was computed from spring snow cover duration information at Canadian climate stations north of 60ºN as follows: the number of days with snow depth greater than or equal to 2 cm in the second half of the hydrological year (January–July), converted to a standardized anomaly with respect to a 1961–90 reference period. Individual station anomalies were averaged over the area north of 60ºN, and the areal average standardized with respect to the 1968–98 period. The inverse series is plotted in Fig. 2b to facilitate the visual comparison with air temperature. Spring snow cover was significantly correlated (r = -0.78, $p<0.01$) with the Jones temperature series (not shown).

c *Open Water Extent* (Fig. 2c)
Homogenized time series of maximum open water (as a percentage) in the QEI inter-island channels and north-eastern Baffin Bay, calculated from combining records from the CIS weekly chart-based digitized database (1968–98) with the manually extracted record from the Polar Continental Shelf Project sea-ice atlas (1961–78). This parameter is also the inverse of minimum sea-ice extent. The values are standardized with respect to the 1968–98 period. The values for 1999 and 2000 are preliminary.

d *Icecap Melt* (Fig. 2d)
Summer melt calculated by taking the inverse of summer mass balance values for the Devon Ice Cap from the 1999 update of Koerner (1997) and standardized to the 1968–98 period.

e *Lake Freeze-up* (Fig. 2e)
This series was derived from a composite of in situ observations of freeze-up at Back Bay (1956–96) and observed freeze-up for Great Slave Lake (1988–99) from SSM/I passive microwave satellite data. The difference in the two series for the 1988–96 period of overlapping data was used to adjust the Back Bay values which froze an average of 34 days earlier than the entire lake. The combined data series was converted to standardized anomalies with respect to the 1968–98 period.

f *Ground Temperature* (Fig. 2f)
This series was derived from daily soil temperature measurements (morning observations) made at a depth of 100 cm at the Resolute Airport climate station. These data are part of the Canadian soil temperature network established in 1958 by the Meteorological Service of Canada. Currently only two Arctic stations (Resolute and Clyde) remain in the network which had declined to 28 stations from a maximum of 68 at the end of 1980. The July monthly mean was computed and converted to standardized anomalies with respect to the 1968–98 period. This particular month has the most complete data.

References

AGNEW, T. and A. SILIS. 1995. Spring-season climate variability in the Canadian Arctic Islands. *Ann. Glaciol.* **21**: 330–336.

———; B.T. ALT, R. DE ABREU and S. JEFFERS. 2001. The loss of decade old sea ice plugs in the Canadian Arctic Islands. *In:* Extended abstracts: Polar Meteorology and Oceanography Conference, Am. Meteorol. Soc., San Diego, CA. 14–18 May, 376 pp.

ALT, B.T. 1987. Developing synoptic analogs for extreme mass balance conditions on Queen Elizabeth Island ice caps. *J. Clim. Appl. Meteorol.* **26**: 1605–1623.

———. 2001. Synoptic climate of 1998 in context and in detail in the Queen Elizabeth Islands. *In*: Chapter II:1 The state of the Arctic cryosphere during the extreme warm summer of 1998: documenting cryospheric variability in the Canadian Arctic. CCAF summer 1998 Project Team, CCAF Final Report. Available at http://www.socc.ca, 89 pp.

ATKINSON, D.E. and K. GAJEWSKI. 2002. High-resolution estimation of surface air temperature in the Canadian Arctic Archipelago. *J. Clim.* **15**: 3601–3614.

BAMZAI, A.S. 2003. Relationship between snow cover variability and Arctic oscillation index on a hierarchy of time scales. *Int. J. Climatol.* **23**: 131–142.

BARRY, R.G. and J.A. MASLANIK. 1993. Monitoring lake freeze-up/break-up as a climatic index. *In:* Proc. Snow Watch '92, R. G. Barry, B. E. Goodison and E. F. LeDrew (Eds), World Data Center A for Glaciology, Glaciological Data Report GD-25, pp 66–79.

BOON, S.; M. SHARP and P. NIENOW. 2003. Impact of an extreme melt event on the runoff and hydrology of a high Arctic glacier. *Hydrol. Process.* **17**: 1051–1072.

BOURGEOIS, J.C.; R.M.KOERNER, K. GAJEWSKI and D.A. FISHER. 2000. A holocene ice-core pollen record from Ellesmere Island, Nunavut, Canada. *Quat. Res.* **54**: 275–283.

———; K. GAJEWSKI and R.M. KOERNER. 2001. Spatial patterns of pollen deposition in Arctic snow. *J. Geophys. Res.* **106** (D6): 5255–5565.

BROWN, J.; K.M. HINKEL and F.E. NELSON (Eds). 2000. The Circumpolar Active Layer Monitoring (CALM) program: Research designs and initial results. Special Issue, *Polar Geog.* **24** (3): 165–258.

BROWN, R.D. 2000. Northern hemisphere snow cover variability and change, 1915–1997. *J. Clim.* **13**: 2339–2355.

——— and D. O'NEILL. 2002. National plan for cryospheric monitoring – a Canadian contribution to the Global Climate Observing Program. Meteorological Service of Canada, Climate Processes and Earth Observations Division, Downsview, Ontario, M3H 5T4, 85 pp. (available at http:www.socc.ca/gcos_cry_plan_canada.pdf)

CCAF SUMMER 1998 PROJECT TEAM. 2001. The state of the Arctic cryosphere during the extreme warm summer of 1998: documenting cryospheric variability in the Canadian Arctic. CCAF Final Report. Available from http://www.socc.ca .

CHAPMAN W. L. and J. E. WALSH. 1993. Recent variations of sea ice and air temperature in high latitudes. *Bull. Am. Meteorol. Soc*. **74**: 33–47.

COGLEY, J.G.; W.P. ADAMS, M.A. ECCLESTONE, F. JUNG-ROTHENHAUSLER and C.S.L. OMMANNEY. 1995. Mass balance of White Glacier, Axel Heiberg Island, N.W.T, Canada, 1960–91 - a reassessment and discussion. National Hydrology Research Institute Science Report No. 6, 168 pp.

COMISO, J. 1990. DMSP SSM/I daily polar gridded sea ice concentrations, June to September 2001. J. Maslanik and J. Stroeve (Eds). National Snow and Ice Data Center, Boulder, CO. Digital media updated to current year.

———. 2002. A rapidly declining perennial sea ice cover in the Arctic. *Geophys. Res. Lett.* **29**: 1956, doi: 1029/2002GL015650.

EDLUND, S.A.; B.T. ALT and K.L. YOUNG. 1989. Interaction of climate, vegetation, and soil hydrology at Hot Weather Creek, Fosheim Peninsula, Ellesmere Island, Northwest Territories. Current Research, Part D, Geological Survey of Canada, Paper 89-1D, pp.125–133.

ENVIRONMENT CANADA, 2002. Winter 1998 temperature and precipitation in historical perspective. Climate Trends and Variations Bulletin for Canada. (Available from http://www.msc-smc.ec.gc.ca/ccrm/bulletin/winter98/index_e.html.)

FISHER, D.A. and R.M. KOERNER. 2003. Holocene ice-core climate history: a multi-variable approach. In: *Global Change in the Holocene,* A. Mackay, R. Batterbee, J. Birks, F. Oldfield (Eds), Arnold, London, pp. 281–293.

FLATO, G.M. and R.D. BROWN. 1996. Variability and climate sensitivity of landfast Arctic sea ice. *J. Geophys. Res.* **101**(C10): 25,767–25,777.

GAJEWSKI K. and D.E. ATKINSON. 2003. Climatic change in northern Canada. *Environment. Rev.* **11**: 69–102.

GLOERSEN, P. 1995. Modulation of hemispheric sea-ice cover by ENSO events. *Nature,* **373**: 503–506.

GOODISON, B.E.; R.D. BROWN and R.G. CRANE (coordinators) with contributions from R. ALLEY, R. BALES, D. BARBER, R.G. BARRY, R.D. BROWN, C. BENTLEY, T. CARROL, D. CLINE, C.R. DUGUAY, G.M. FLATO, D.K. HALL, R. HARRINGTON, J. KARGEL, H. KIEFFER, S. MUNRO, C. PARKINSON, B. RAUP, A. ROTHROCK and M. SHARP. 1999. Cryospheric Systems. *In:* EOS Science Plan – Chapter 6, National Aeronautics and Space Administration. pp. 261–307.

GROISMAN, P. YA; T. R. KARL and R.W. KNIGHT. 1994. Observed impact of snow cover on the heat balance and the rise of continental spring temperatures. *Science*, **263**: 198–200.

GRUMET, N.S.; C.P. WAKE, P.A. MAYEWSKI, G.A. ZIELINSKI, S.I. WHITLOW, R.M. KOERNER, D.A. FISHER and J.M. WOOLLETT. 2001. Variability of sea-ice extent in Baffin Bay over the last millennium. *Climatic Change*, **49**: 129–145.

HARRIS, C. and A.G. LEWKOWICZ. 2000. An analysis of the stability of thawing slopes, Ellesmere Island, Nunavut, Canada. *Canadian Geotech. J.* **37**(2): 465–478.

IPCC. 2001. *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change.* Houghton, J.T., Y. Ding, D.J. Griggs, M. Noguer, P.J. van der Linden, X. Dai, K. Maskell and C.A. Johnson (Eds). Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 881 pp.

JEFFERIES, M.O.; G.J. REYNOLDS and J.M. MILLER. 1992. First Landsat multispectral scanner images of the Canadian Arctic north of 80ºN. *Polar Rec.* **28**: 1–6.

JEFFERS, S.; T.A. AGNEW, B.T. ALT, R. DE ABREU and S. MCCOURT. 2001. Investigating the anomalous sea ice conditions in the Canadian High Arctic (Queen Elizabeth Islands) during the summer of 1998. *Ann. Glaciol.* **33**: 507–512.

JOHANNESSEN, O.M.; E.V. SHALINA and M.W. MILES. 1999. Satellite evidence for an Arctic sea ice cover in transformation. *Science*, **286**: 1937–1939.

JONES, P.D. and A. MOBERG. 2003. Hemispheric and large-scale surface air temperature variations: An extensive revision and an update to 2001. *J. Clim.* **16**: 206–223.

KALNAY, E.; M. KANAMITSU, R. KISTLER, W. COLLINS, D. DEAVEN, L. GANDIN, M. IREDELL, S. SAHA, G. WHITE, J. WOLLEN, Y. ZHU, A. LEETMAA, B. REYNOLDS, M. CHELLIAH, W. EBISUZAKI, W. HIGGINS, J. JANOWIAK, K.C. MO, C. ROPELEWSKI and J. WANG. 1996. The NCEP/NCAR reanalysis 40-year Project. *Bull. Am. Meteorol. Soc.* **77**: 437–471.

KOERNER, R.M. 1997. Some comments on climatic reconstructions from ice cores drilled in areas of high melt. *J. Glaciol.* **43** (143): 90–97.

——— and B.T. ALT. 2001. Temporal and spatial nature of change in the circumpolar Arctic. *In:* Proc. Second Wadati Conference on Global Change and Polar Climate, 7–9 March 2001, Tsukuba, Japan, pp. 51–53.

LANZANTE, J.R.; S.A. KLEIN and D.J. SEIDEL. 2003. Temporal homogenization of monthly radiosonde temperature data. Part II: Trends, sensitivities, and MSU comparison. *J. Clim.* **16**: 241–262.

LEWKOWICZ, A.G. 1990. Morphology, frequency and magnitude of active-layer detachment slides, Fosheim Peninsula, Ellesmere Island, N.W.T. *In:* Permafrost - Canada: Proceedings of the Fifth Canadian Permafrost Conference. 2–5 August 1998, Université Laval, Québec City, pp. 111–118.

———. 1992. Factors influencing the distribution and initiation of active-layer detachment slides on Ellesmere Island, Arctic Canada. In: *Periglacial Geomorphology, Proc. of the 22nd Annual Binghamton Symposium in Geomorphology,* J.C. Dixon and Abrahams, A.D. (Eds). John Wiley and Sons, Chichester, pp. 223–250.

MAGNUSON, J.J.; D.M. ROBERTSON, B.J. BENSON, R.H. WYNNE, D.M. LIVINGSTONE, T. ARAI, R.A. ASSEL, R.G. BARRY, V. CARD, E. KUUSISTO, N.G. GRANIN, T.D. PROWSE, K.M. STEWART and V.S. VUGLINSKI. 2000. Historical trends in lake and river ice cover in the Northern Hemisphere. *Science*, **289**: 1743–1746.

MASLANIK, J.; M. SERREZE and R. BARRY. 1996. Recent decreases in Arctic summer ice cover and linkages to atmospheric circulation anomalies. *Geophys. Res. Lett.* **23**: 1677–1680.

———;——— and T. AGNEW. 1999. Record reduction in western Arctic sea ice cover in 1998: Characteristics and relationships to atmospheric circulation. *Geophys. Res. Lett.* **26**: 1905–1908.

MCNUTT, S.L. and J.E. OVERLAND. 2003. Spatial hierarchy in Arctic sea ice dynamics. *Tellus Series A - Dyn. Meteorol. Oceanogr.* **55**: 181–191.

MELLING, H. 2002. Sea ice of the northern Canadian Arctic Archipelago. *J. Geophys. Res. - Oceans*, **107** (C11): 3181, doi:10.1029/2001JC001102.

MILEWSKA, E. 2004. Baseline cloudiness trends in Canada 1953–2002. ATMOSPHERE-OCEAN, **42** (4): 267–280.

NIEBAUER, H.J. 1999. The 1997–98 El Niño in the Bering Sea as compared with previous ENSO events and the "regime shift" of the late 1970s. *In:* Proceedings of the 1998 Science Board Symposium on the impacts of the 1997/98 El Nino event on the North Pacific Ocean and its marginal seas. October 1998, North Pacific Marine Science Organization, 131 pp.

ONO, N. 1996. ENSO-like periodicities in the Arctic Cryosphere. *In:* Proceedings of the International Symposium on Environmental Research in the Arctic, 19–21 July 1995, Tokyo, 398 pp.

OSTERMEIER, G.M. and J.M. WALLACE. 2003. Trends in the North Atlantic Oscillation-Northern Hemisphere annular mode during the twentieth century. *J. Clim.* **16**: 336–341.

PAPINEAU, J.M. 2001. Wintertime temperature anomalies in Alaska correlated with ENSO and PDO. *Int. J. Climatol.* **21**: 1577–1592

PARKINSON, C.L. 1992. Spatial patterns of increases and decreases in the length of the sea ice season. *J. Geophys. Res.* **97**(C9): 14,377–14,388.

———; D.J. CAVALIERI, P. GLOERSEN, H.J. ZWALLY and J.C. COMISO. 1999. Arctic sea ice extents, areas, and trends, 1978–1996. *J. Geophys. Res.* **104**(C9): 20837–20856.

PHILLIPS. D. (Ed.). 1995. The Global Climate System Review. Climate System Monitoring; June 1991 - November 1993. World Meteorological Organization, No. 819, 150 pp.

POLYAKOV, I.V.; G.V. ALEKSEEV, R.V. BEKRYAEV, U. BHATT, R.L. COLONY, M.A. JOHNSON, V.P. KARKLIN, A.P. MAKSHTAS, D. WALSH and A.V. YULIN. 2002. Observationally based assessment of polar amplification of global warming. *Geophys. Res. Lett.* **29**: 1878, doi: 10.1029/2001GL011111.

PROSHUTINSKY, A.Y. and M.A. JOHNSON. 1997. Two circulation regimes of the wind-driven Arctic Ocean. *J. Geophys. Res.* **102** (C6): 12493–12514.

———; R.H. BOURKE and F.A. MCLAUGHLIN. 2002. The role of the Beaufort Gyre in Arctic climate variability: Seasonal to decadal climate scales. *Geophys. Res. Lett.* **29**: 2100, doi:10.1029/2002GL015847.

RIGOR, I.G.; J.M. WALLACE and R.L. COLONY. 2001. Response of sea ice to the Arctic Oscillation. *J. Clim.* **15**: 2648–2663.

ROMANOVSKY, V.; M. BURGESS, S. SMITH, K. YOSHIKAWA and J. RROWN. 2002. Permafrost temperature records: Indicators of climate change. *EOS, Trans. Am. Geophys. Union*. **83** (50): 589–594.

ROUSE, W.R.; E.M. BLYTH, R.W. CRAWFORD, J.R. GYAKUM, J.R. JANOWICZ, B. KOCHTUBAJDA, H.G. LEIGHTON, P. MARSH, L. MARTZ, A. PIETRONIRO, H. RITCHIE, W.M. SCHERTZER, E.D. SOULIS, R.E. STEWART, G.S. STRONG and M.K. WOO. 2003a. Energy and water cycles in a high-latitude, north-flowing river system - Summary of results from the Mackenzie GEWEX Study - Phase I. *Bull. Am. Meteorol. Soc.* **84**: 73–87.

———; C. M. OSWALD, J. BINYAMIN, P. D. BLANKEN, W.M. SCHERTZER and C. SPENCE. 2003b. Interannual and seasonal variability of the surface energy balance and temperature of central Great Slave Lake. *J. Hydrometeorol.* **4**: 720–730.

———;———,———, C. SPENCE, W.M. SCHERTZER, P.D. BLANKEN, N. BUSSIÈRES and C.R. DUGUAY. 2005. The role of northern lakes in a regional energy balance. *J. Hydrometeorol.* **6**(3): 291–305.

SCHWEIGER, A. J. 2004. Changes in seasonal cloud cover over the Arctic seas from satellite and surface observations. *Geophys. Res. Lett.* **31**: L12207, doi:10.1029/2004GL020067.

SERREZE, M.; J. WALSH, F. CHAPIN, T. OSTERKAMP, M. DYURGEROV, V. ROMANOVSKY, W. OECHEL, J. MORRISON, T. ZHANG and R.G. BARRY. 2000. Observational evidence of recent changes in the northern high-latitude environment. *Clim. Change*, **46**: 159–207.

SMITH, S.L. and M.M. BURGESS. 1999. Mapping the sensitivity of Canadian permafrost to climate warming. *In:* Interactions between the cryosphere, climate and greenhouse gases; Proceedings of IUGG 99 Symposium HS2, Birmingham. M. Tranter, R. Armstrong, E. Brun, G. Jones, M. Sharp and M. Williams (Eds), July 1999. IAHS Publication no. 256, pp. 71–80.

———;——— and F.M. NIXON. 2001a. Response of active-layer and permafrost temperatures to warming during 1998 in the Mackenzie Delta, Northwest Territories and at Canadian Forces Station Alert and Baker Lake, Nunavut; Geological Survey of Canada, Current Research 2001-E5, 8 pp.

———;——— and J.A. HEGINBOTTOM. 2001b. Permafrost in Canada, a challenge to northern development. *In:* A synthesis of Geological Hazards in Canada, G.R. Brooks (Ed.), Geological Survey of Canada, Bulletin 548, pp. 241–264.

———;——— and A.E. TAYLOR. 2003. High Arctic permafrost observatory at Alert, Nunavut – analysis of a 23-year data set. *In:* Permafrost: 8th International Conference, Zurich, M. Phillips, S.M. Springman, L. Arenson and A.A. Balkema (Eds), 21–25 July 2003, Lisse, Switzerland.

TRENBERTH, K.E. and T.J. HOAR. 1996. The 1990–95 El Nino-Southern Oscillation event: Longest on record. *Geophys. Res. Lett.* **23**: 57–60.

VINNIKOV, K.; A. ROBOCK, R.J. STOUFFER, J.E. WALSH, C.L. PARKINSON, D.J. CAVALIERI, J.F.B. MITCHELL, D. GARRETT and V.F. ZAKHAROV. 1999. Global warming and Northern Hemisphere sea ice extent. *Science,* **286**: 1934–1937.

WALKER, A.E.; J.L. SOKOL and M.R. DAVEY. 1997. Use of SSM/I 85 GHz passive microwave data for investigating Great Slave Lake ice freeze-up/breakup variations. *In:* Proc. GER'97 (International Symposium: Geomatics in the Era of Radarsat), 24–30 May 1997, Ottawa, Ontario (CD-ROM).

WOLFE, S.A.; E. KOTLER and F.M. NIXON. 2000. Recent warming impacts in the Mackenzie Delta, Northwest Territories, and northern Yukon Territory coastal areas; Geological Survey of Canada, Current Research 2000-B1, 9 pp.

YOUNG, K.L. and A.G. LEWKOWICZ. 1990. Surface energy balance of a perennial snowbank, Melville Island, Northwest Territories, Canada. *Arctic Alpine Res.* **22** *(*3): 290–301.

——— and M-K. WOO. 2000. Hydrological response of a patchy high Arctic wetland. *Nordic Hydrol*. **31**: 317–238.

——— and ———. 2003. Thermo-hydrological responses to an exceptionally warm, dry summer in a High Arctic environment. *Nordic Hydrol.* **34**: 51–70.

ZHANG, X.; L.A. VINCENT, W.D. HOGG and AIN NIITSOO. 2000. Temperature and precipitation trends in Canada during the 20th Century. ATMOSPHERE-OCEAN, **38**(3): 395–429.