OBSERVATIONAL EVIDENCE OF RECENT CHANGE IN THE NORTHERN HIGH-LATITUDE ENVIRONMENT

M. C. SERREZE[1], J. E. WALSH[2], F. S.CHAPIN III[3], T. OSTERKAMP[3], M. DYURGEROV[4], V. ROMANOVSKY[3], W. C. OECHEL[5], J. MORISON[6], T. ZHANG[1] and R. G. BARRY[1]

[1]*Cooperative Institute for Research in Environmental Sciences, Division of Cryospheric and Polar Processes, Campus Box 449, University of Colorado, Boulder, CO 80309-0449, U.S.A.*
E-mail: serreze@kryos.colorado.edu
[2]*Department of Atmospheric Sciences, University of Illinois, Urbana-Champaign, IL, U.S.A.*
[3]*Geophysical Institute, University of Alaska, Fairbanks, AK, U.S.A.*
[4]*Institute of Arctic and Alpine Research, University of Colorado, Boulder, CO, U.S.A.*
[5]*Global Change Research Group, San Diego State University, San Diego, CA, U.S.A.*
[6]*Applied Physics Laboratory, University of Washington, Seattle, WA, U.S.A.*

Abstract. Studies from a variety of disciplines document recent change in the northern high-latitude environment. Prompted by predictions of an amplified response of the Arctic to enhanced greenhouse forcing, we present a synthesis of these observations. Pronounced winter and spring warming over northern continents since about 1970 is partly compensated by cooling over the northern North Atlantic. Warming is also evident over the central Arctic Ocean. There is a downward tendency in sea ice extent, attended by warming and increased areal extent of the Arctic Ocean's Atlantic layer. Negative snow cover anomalies have dominated over both continents since the late 1980s and terrestrial precipitation has increased since 1900. Small Arctic glaciers have exhibited generally negative mass balances. While permafrost has warmed in Alaska and Russia, it has cooled in eastern Canada. There is evidence of increased plant growth, attended by greater shrub abundance and northward migration of the tree line. Evidence also suggests that the tundra has changed from a net sink to a net source of atmospheric carbon dioxide.

Taken together, these results paint a reasonably coherent picture of change, but their interpretation as signals of enhanced greenhouse warming is open to debate. Many of the environmental records are either short, are of uncertain quality, or provide limited spatial coverage. The recent high-latitude warming is also no larger than the interdecadal temperature range during this century. Nevertheless, the general patterns of change broadly agree with model predictions. Roughly half of the pronounced recent rise in Northern Hemisphere winter temperatures reflects shifts in atmospheric circulation. However, such changes are not inconsistent with anthropogenic forcing and include generally positive phases of the North Atlantic and Arctic Oscillations and extratropical responses to the El-Niño Southern Oscillation. An anthropogenic effect is also suggested from interpretation of the paleoclimate record, which indicates that the 20th century Arctic is the warmest of the past 400 years.

1. Introduction

The Arctic has attained a prominent role in scientific debate regarding global climate change. General circulation models (GCMs) predict that the effects of

Climatic Change **46:** 159–207, 2000.

anthropogenic greenhouse warming will be amplified in the northern high latitudes due to feedbacks in which variations in snow and sea ice extent, the stability of the lower troposphere and thawing of permafrost play key roles. Although Arctic warming is somewhat diminished when anthropogenic change experiments include sulfate aerosol effects and coupling to a deep ocean and regional patterns of warming differ among simulations, polar amplification in the Northern Hemisphere remains a characteristic feature of model predictions. Projected warming is greatest for late autumn and winter, largely because of the delayed onset of sea ice and snow cover. Retreat of snow cover and sea ice is accompanied by increased winter precipitation (Nicholls et al., 1996). If models are correct in their depiction of amplified warming in the Arctic, the observed buildup of greenhouse gas concentrations (an equivalent increase of carbon dioxide by about 50% since the mid 18th century) should by now arguably be producing detectable climate signals.

However, change detection is hampered by deficiencies in data sets. Gridded atmospheric fields are available from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis since the late 1950s (Kalnay et al., 1996) and certain atmospheric indices can be constructed for longer periods. Terrestrial records for key Arctic monitoring variables such as air temperature and precipitation (Groisman et al., 1991; Eischeid et al., 1995), glacier mass balance (Dowdeswell, 1997; Dyurgerov and Meier, 1997) and permafrost conditions (Osterkamp and Romanovsky, 1996) have problems of spatial sampling and are often short or discontinuous. Cloud cover analyses are largely restricted to climatologies based on surface observations (Warren et al., 1988) or fairly short satellite-derived time series of uncertain quality (Rossow and Schiffer, 1991). Gridded fields of sea ice extent derived from satellite passive microwave imagery or by combining different imagery types with aircraft and ship reports are available from the early 1970s onwards (Parkinson and Cavalieri, 1989; Gloersen and Campbell, 1991; Chapman and Walsh, 1993; Maslanik et al., 1996). Reliable data sets of Arctic Ocean air temperatures have only recently been compiled, based on Russian 'North Pole' (NP) drifting station records and blending the NP data with buoy measurements from the International Arctic Buoy Program (IABP) and coastal station observations (Martin et al., 1997; Martin and Munoz, 1997). Precipitation time series for the Arctic Ocean are largely based on NP records (Colony et al., 1997).

A principal goal of the National Science Foundation (NSF) Arctic System Science (ARCSS) program is to advance the scientific basis for predicting Arctic environmental change on decade to century time scales. Achieving this goal will require improved data bases and the development of models to quantify interactions between the atmospheric, terrestrial, oceanic and human components of the Arctic system and linkages with global processes. These requirements are being addressed through several linked programs and sub-programs of ARCSS including Land Atmosphere Ice Interactions (LAII), Ocean Atmosphere Interactions (OAI), Paleoclimates of Arctic Lakes and Estuaries (PALE), Shelf-Basin Interactions (SBI) and the Human Dimensions of the Arctic System (HARC). Achieving the goals of ARCSS also requires synthesis of existing research. As a contribution to ARCSS, the present paper reviews observational evidence of change in the Arctic environment to assess the extent to which observations are consistent with climate model predictions. The evidence reviewed here has been primarily delivered by efforts through National Weather Services of northern countries, NSF, the National Aeronautics and Space Administration (NASA), the National Oceanic and Atmospheric Administration (NOAA), the U.S. Navy and the Russian North Pole Program.

2. Atmosphere and Surface Climate

2.1. SURFACE AIR TEMPERATURE

Surface air temperature is the most commonly cited climate change variable. It is particularly useful as it integrates changes in the surface energy budget and atmospheric circulation. From analysis of records examined as part of the Intergovernmental Panel on Climate Change (IPCC) assessment (Nicholls et al., 1996), global mean surface air temperatures have risen by about 0.3 °C to 0.6 °C since the late 19th century and by 0.2 °C to 0.3 °C over the past 40 years for which data are most reliable (0.050–0.075 °C per decade). Both hemispheres have participated in this warming. The largest temperature increases in recent decades have occurred over Northern Hemisphere land areas from about 40–70° N. On the basis of proxy sources (e.g., tree rings and varves) that are primarily indicators of summer conditions, Overpeck et al. (1997) conclude that Arctic temperatures in the 20th century are the highest in the past 400 years.

Figure 1 shows the spatial pattern of annual mean surface air temperature trends for the Northern Hemisphere north of 40° N over the period 1966–1995. Results represent an update of the Chapman and Walsh (1993) analysis, based on data from Jones (1994). Temperatures during this period have increased markedly over the Eurasian and northwest North American land masses. Locally trends exceed 0.5 °C per decade. This warming remains when stations suspected of having urban influences are removed from the data set (Jones, 1994). Over the ocean basins, temperature changes are generally smaller or negative. Pronounced cooling characterizes the western subpolar north Atlantic and extends into land areas over eastern Canada and southern Greenland. As illustrated in the corresponding maps for individual seasons (Figure 2), the annual results are dominated by trends for winter and to a lesser extent spring. Spatial trend patterns for summer and autumn are weaker, with autumn showing small negative trends over northern North America and Europe.

Temperature records for 1900–1995 expressed as means for the 55–85° N zonal band (based primarily on land stations) (Figure 3) place the results from Figures 1 and 2 into a longer-term perspective. It is apparent that our interpretation of temperature trends changes substantially if decades prior to 1970 are included. Annual

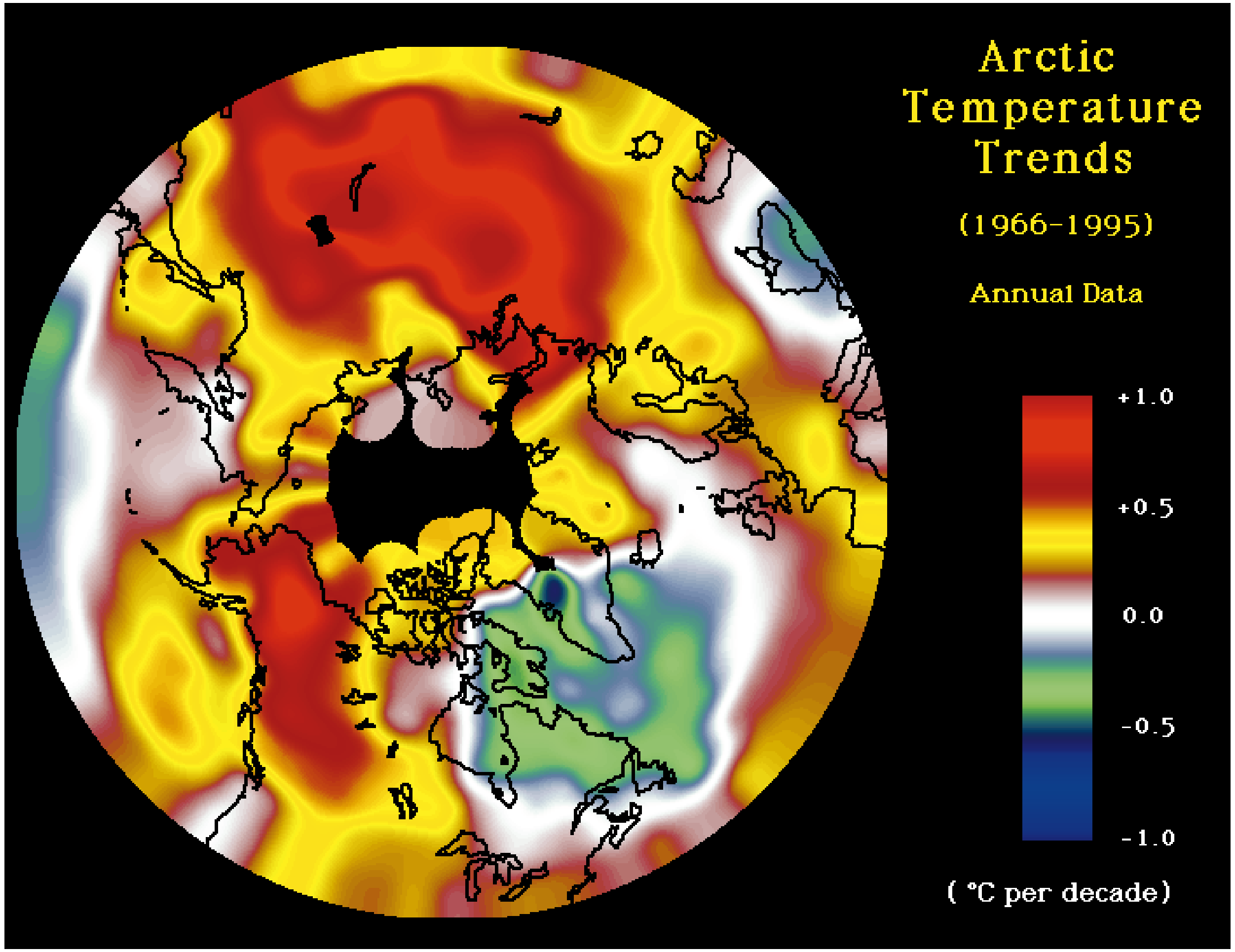

Figure 1. Trends in mean annual surface air temperatures in °C per decade north of 40° N for the period 1966–1995. Areas with insufficient data are shown in black (updated from Chapman and Walsh, 1993).

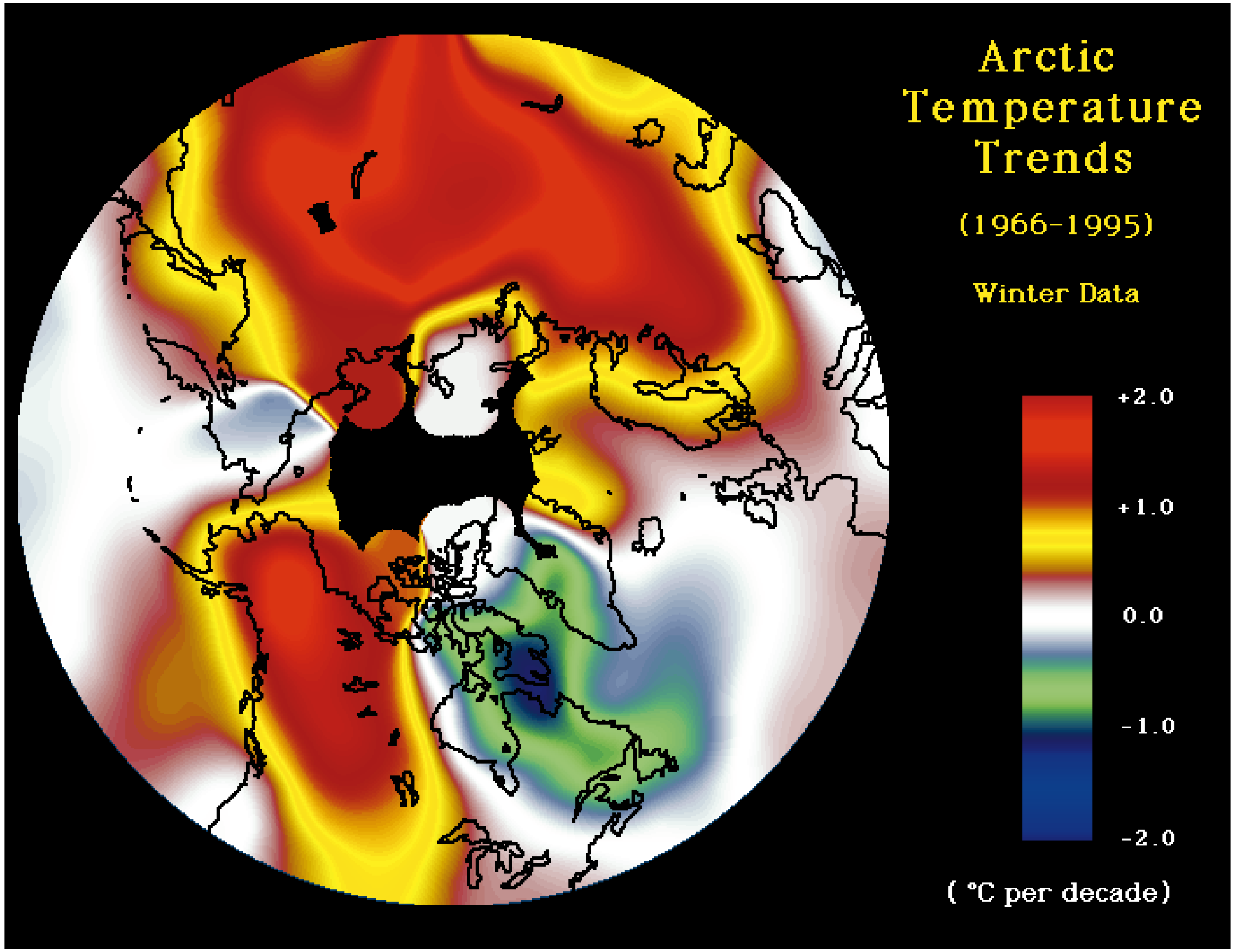

Figure 2a. Winter.

Figure 2. Trends in mean surface air temperatures (°C per decade) north of 40° N for the period 1966–1995 for: (a) winter; (b) spring; (c) summer; (d) autumn. Areas with insufficient data are shown in black (updated from Chapman and Walsh, 1993).

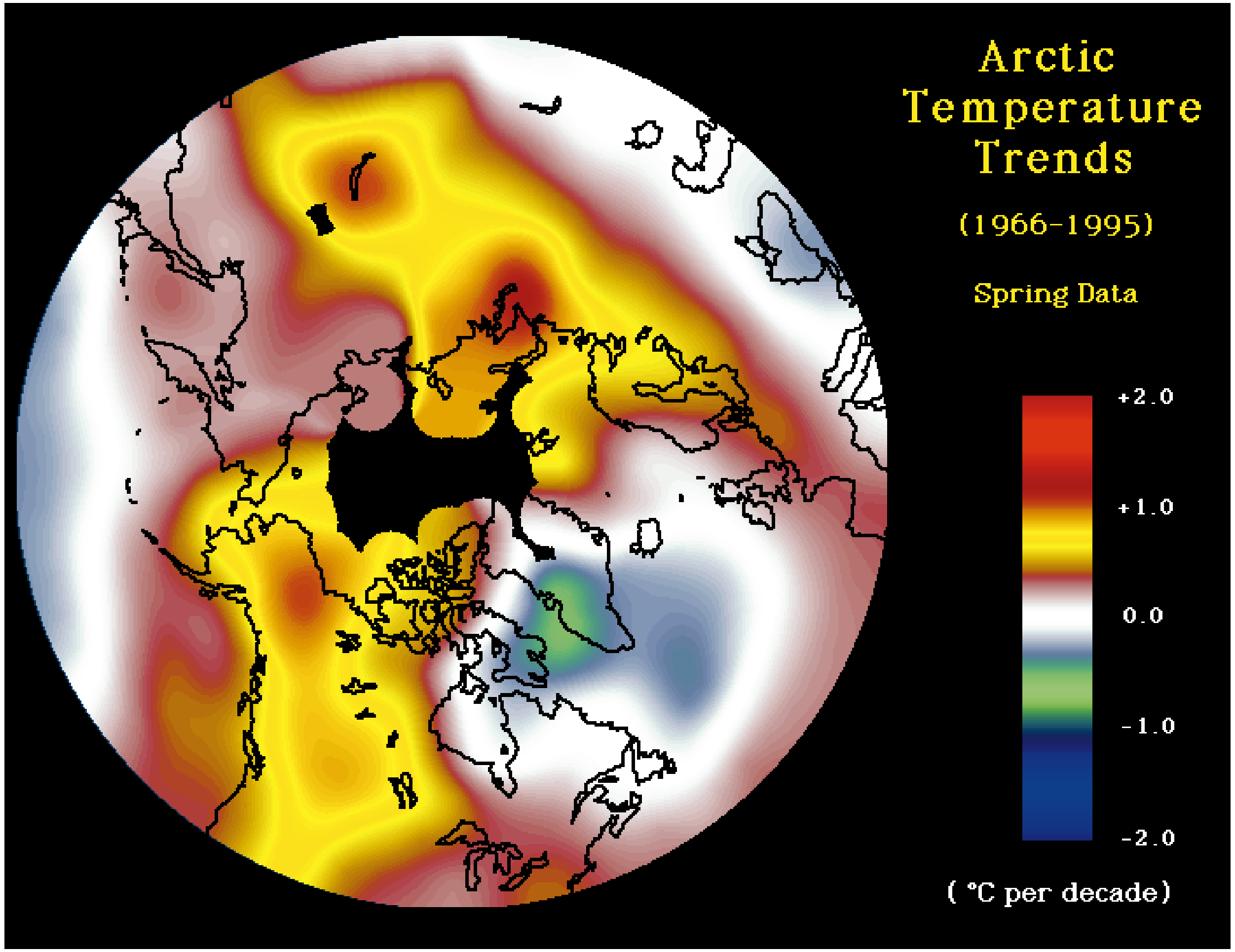

Figure 2b. Spring.

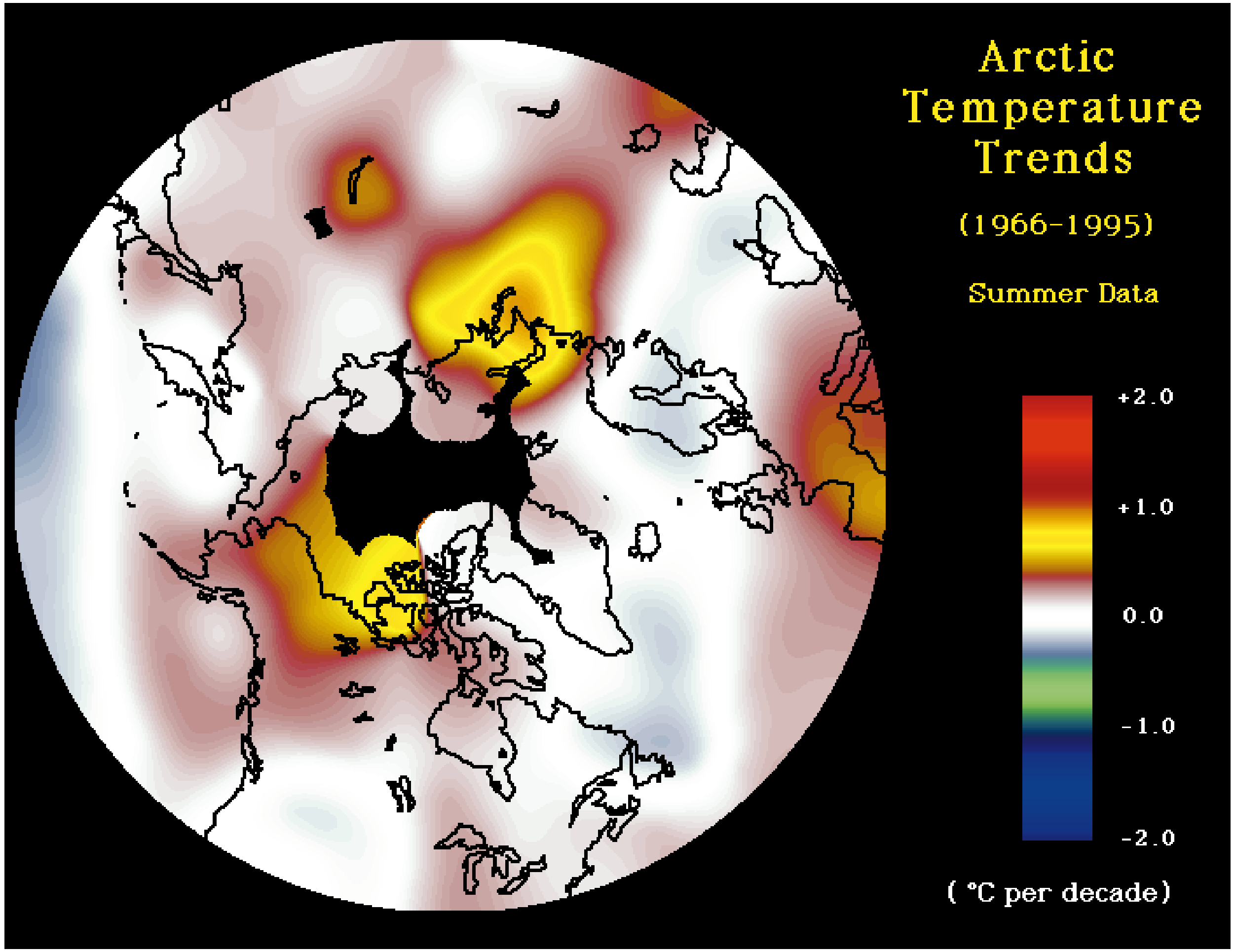

Figure 2c. Summer.

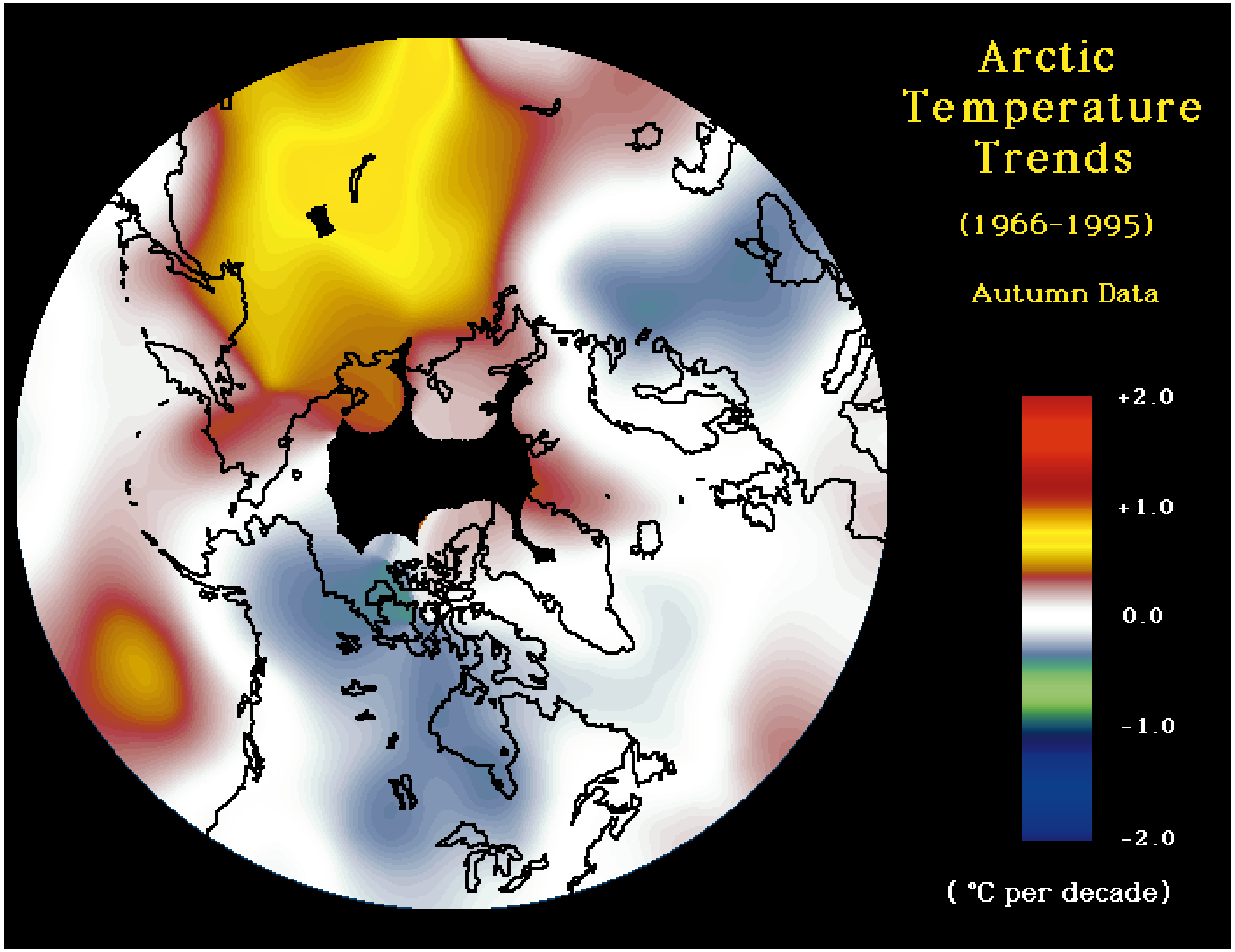

Figure 2d. Autumn.

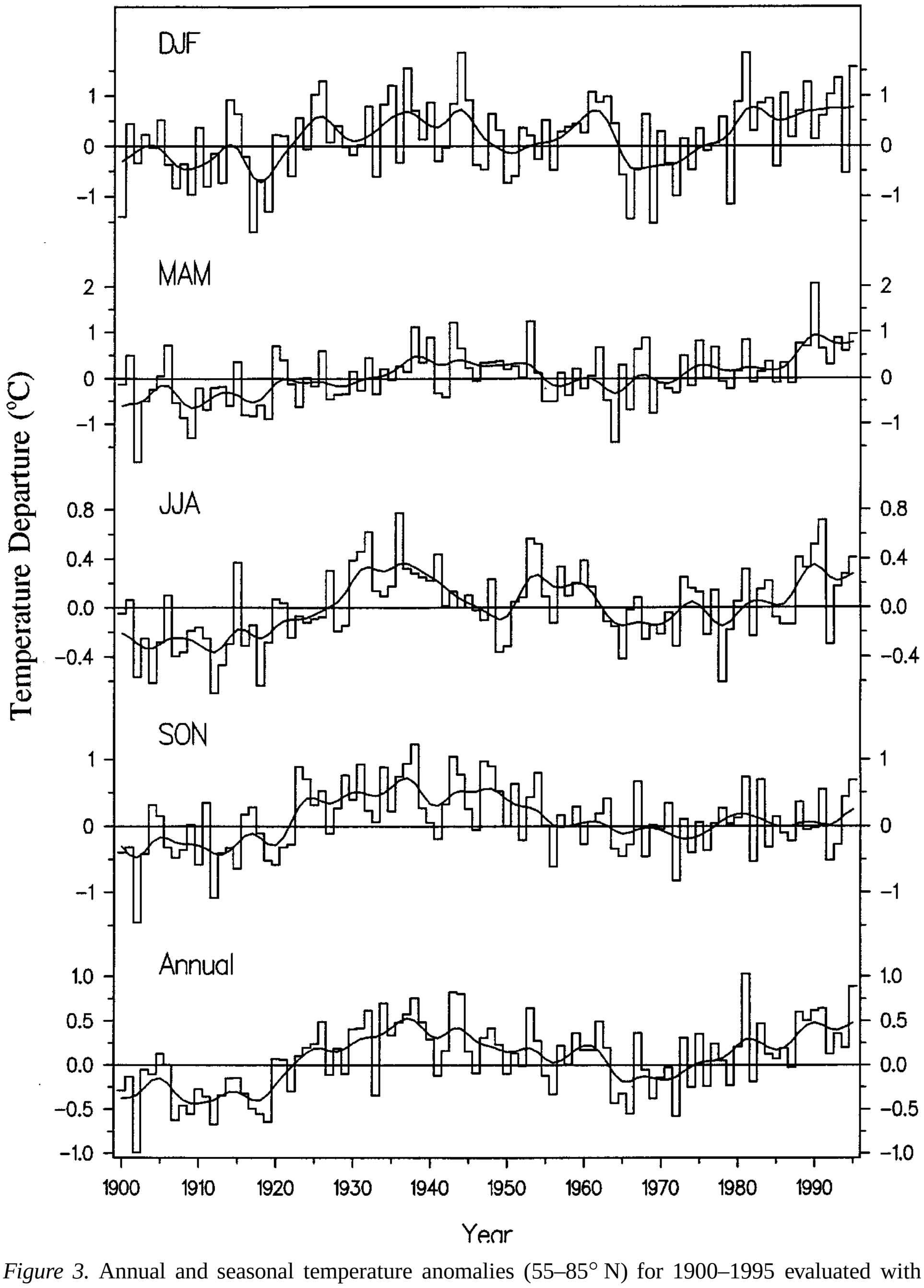

Figure 3. Annual and seasonal temperature anomalies (55–85° N) for 1900–1995 evaluated with respect to 1951–1980 means (°C). The smoothed line represents results from a nine-point low-pass filter. Results are based on updates to the Eischeid et al. (1995) data set.

mean temperatures fell during the period 1940–1970. While recognizing sampling problems in the early part of the record, it appears that annual temperatures from 1920–1940 rose even more markedly than during the post 1970s period.

Recent warming, however, is not in doubt and appears to extend into the central Arctic Ocean. By combining all Russian NP drifting station records from 1961–1990, Martin et al. (1997) find statistically significant increases in air temperature during May and June respectively of 0.89 °C and 0.43 °C per decade, as well as significant increases for summer as a whole. These results are seemingly contradicted by Kahl et al. (1993), who examined seasonal trends in central Arctic Ocean air temperatures for 1950–1990 using a combination of dropsonde data from the U.S. Ptarmigan weather reconnaissance aircraft and rawinsonde data from the NP drifting stations. They find significant changes only for autumn and winter, when the temperature cools at rates of 1.0 °C and 0.6 °C per decade, respectively. The reasons for this discrepancy are unclear although Martin et al. (1997) argue the possibility of warm biases in the dropsonde measurements (which dominate the early years of the records), related to a time lag of the dropsonde response as it falls through the near-surface boundary layer. We also examined the new gridded POLES 2-meter air temperature data set for 1979–1995, which blends the NP data with IABP drifting buoy and coastal station records. While temperature trends derived from this data set should be viewed cautiously, due both to the tendency for the buoys to overestimate summer temperatures because of radiational heating and the shorter record available, results indicate that over the 17-year record, warming has occurred from January through July (Figure 4).

To place recent warming in the perspective of the past several centuries, Overpeck et al. (1997) attempt to explain variability in their reconstructed 400 year Arctic temperature record (Figure 5) in terms of the relative roles of changes in trace gas loading, irradiance (solar radiation), aerosol loading from volcanic eruptions and atmospheric circulation. They conclude from their statistical analysis that the pronounced Arctic warming between 1820 and 1920 is primarily due to reduced forcing by volcanic aerosols and increasing irradiance. After 1920, both high insolation and low aerosol loading likely continued to influence Arctic climate, but exponentially increasing trace gas concentrations probably played an increasingly dominant role. Based on a reconstructed time series, Lean et al. (1995) conclude that approximately half of the observed Northern Hemisphere warming since 1860 and a third of the warming since about 1970 is attributable to increasing insolation. From a re-working of these data, Overpeck et al. (1997) argue that for the Arctic region, insolation changes have had less of an effect on temperature, but it nevertheless seems that the impacts of solar variability may be larger than previously appreciated.

Conclusions that greenhouse-gas forcing has been a significant player in recent Arctic warming must be viewed cautiously. There is general agreement between climate model predictions and observations in terms of annual mean warming over the past several decades and for maximum warming in northern continental re-

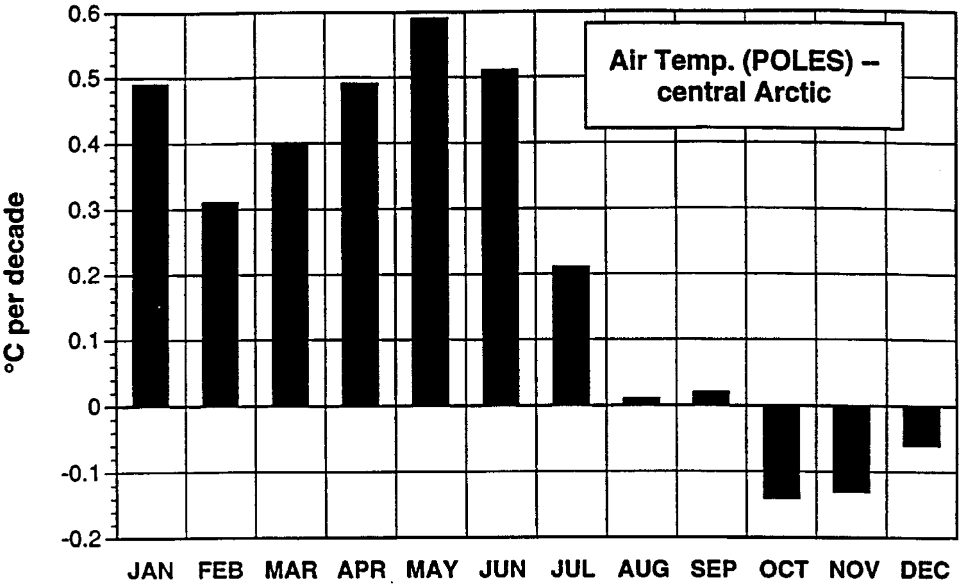

Figure 4. Linear trends (1979–1995) of central Arctic Ocean surface air temperatures in °C per decade based on the POLES data set.

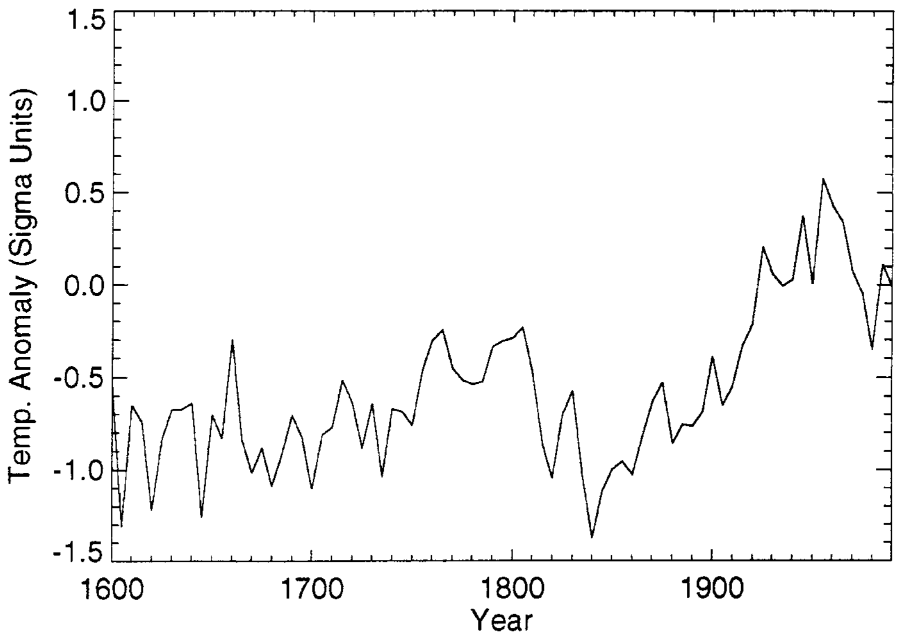

Figure 5. Summer-weighted means of Arctic temperature for 1600–1990 based on proxy records. Results are five-year averages plotted as normalized deviations from observed temperature means for 1901–1960 (data provided by J. Overpeck).

gions. However, discrepancy arises in the seasonality of change. In general, models project the largest warming during late autumn and winter (Kattenberg et al., 1996). By comparison, the observations show maximum winter and spring warming for land, and winter through summer warming over the Arctic Ocean.

Furthermore, the magnitude and spatial structure of Arctic warming vary across models. The Geophysical Fluid Dynamics Laboratory (GFDL) model (Manabe et al., 1992) shows the strongest warming over the Arctic Ocean, particularly the areas of sea ice retreat. The Max Planck Institute (MPI) model shows a similar pattern but the NCAR model shows a much weaker warming and one that is centered over the Pacific subarctic (IPCC, 1992, Figure B4). The United Kingdom Meteorological Office (UKMO) model, on the other hand, shows a warming that is stronger over the subarctic land areas (IPCC, 1992). More recent examples from the Bureau of Meteorology Research Centre (BMRC) and Commonwealth Scientific and Industrial Research Organization (CSIRO) models (Kattenberg et al., 1996) also show a stronger warming over subarctic land areas relative to the subarctic ocean areas.

With regard to these differences, many models suffer from known deficiencies in their parameterizations of high-latitude processes. For example, the models' strong warming in autumn and winter results largely from a delayed freeze-up after an enhancement of ocean heat storage in areas that are newly ice-free during summer (Kattenberg et al., 1996). Yet these changes are highly sensitive to the parameterization of sea ice and its albedo (Meehl and Washington, 1990; Rind et al., 1996). Many GCMs simply prescribe sea ice concentrations (e.g., 90%, 100%) in grid cells in which ice is present; such specifications can grossly oversimplify the spring/summer melt process and the ensuing heat storage in the upper ocean. Cloud feedbacks add to the scatter among models, making it difficult to identify an Arctic 'fingerprint' of greenhouse warming in the model results. This issue is being addressed further in the Coupled Model Intercomparison Project (Meehl et al., 1997).

Wallace et al. (1996) argue that although there is a background temperature change in the Northern Hemisphere that can be viewed as a direct radiative contribution (most clearly manifested in warm-season temperature data), the sharp upward tendency in observed Northern Hemisphere temperatures in recent decades (primarily a reflection of terrestrial warming) is strongly influenced by circulation changes in the cold season. The net effect of these circulation changes is manifested in a tendency towards positive tropospheric thickness anomalies over the high-latitude continents and negative thickness anomalies over the oceans north of 40° N. The patterns in Figures 1 and 2a can be interpreted as the surface signature of this 'cold ocean – warm land' (COWL) pattern (Wallace et al., 1996). It has long been recognized that advection contributes disproportionately to the heat balance of the high latitudes as compared to the low latitudes, especially during the cold season (e.g., Nakamura and Oort, 1988). Consequently, it is reasonable to expect that radiative forcing will be attended by shifts in circulation that may amplify high-latitude temperature changes. Recent model experiments (e.g., Osborn et al., 1999;

Broccoli et al., 1998, Fyfe et al., 1999) support this view. Below, we summarize recent circulation changes and their links with temperature trends.

2.2. ATMOSPHERIC CIRCULATION

Assessing trends and variability in atmospheric circulation, especially for the Arctic, requires recognition that numerical weather prediction models and the amount and quality of available assimilation data have changed considerably. Until 1979, *in situ* observations for the central Arctic Ocean were largely limited to rawinsonde profiles from NP stations. Since that time, incorporation of IABP surface pressure measurements (Colony and Rigor, 1993) has improved the quality of analyzed Arctic fields. Efforts through the NCEP/NCAR reanalysis project provide gridded atmospheric analyses from 1958-present using a ‘frozen’ numerical weather prediction/assimilation system whereby biases due to model changes are eliminated (Kalnay et al., 1996). However, inhomogeneities associated with temporal changes in assimilation data remain (Basist and Chelliah, 1997).

Hurrell (1996) shows that almost half of the wintertime (December–March) temperature variance over the Northern Hemisphere (north of 20° N) since 1935 can be related to the combined effects of circulation variability based on indices of two modes of climate variability: the North Atlantic Oscillation (NAO) (31%) and the Southern Oscillation (SO) (16%). The SO represents the atmospheric component of the El-Niño Southern Oscillation (ENSO) phenomenon, with its index defined from the normalized sea level pressure difference between Tahiti and Darwin. ENSO effects on extratropical circulation, associated with a change towards a more negative SO index over the past two decades, account for part of the winter cooling over the Pacific Basin and warming over northern North America.

The NAO describes a positive relationship between the strength of the Icelandic Low and the Azores High, two of the primary ‘centers of action’ in the Northern Hemisphere general circulation. The positive (negative) phase of the NAO is associated with mutual strengthening (weakening) of the Icelandic Low and Azores High. Under the positive (negative) mode, surface winds tend to be northerly (southerly) over Greenland and eastern Canada, with associated negative (positive) temperature anomalies. Correspondingly, west to southwesterly (northwesterly) winds tend to advect warm, moist (cool and dry) airmasses into northern Europe and Scandinavia. The NAO is best expressed during the cold season. The Icelandic Low and Azores High also tend to be located farther north (south) during the positive (negative) NAO mode (Angell and Korshover, 1974).

While exhibiting considerable interannual variability, the NAO has been in a generally positive phase since about 1970 with several particularly large positive events since about 1980 (Hurrell, 1995, 1996) (Figure 6). The NAO was the strongest in over a century for the winters 1988/1989 to 1994/1995 and there has also been a significant eastward displacement of the Icelandic Low in summer since the 1970s (Machel et al., 1998). It shifted southward of its mean location

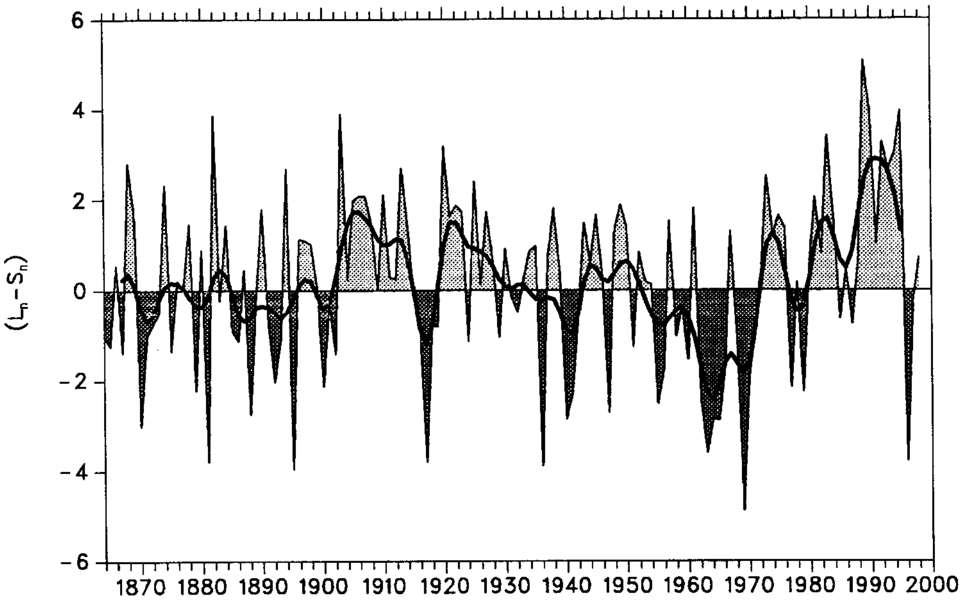

Figure 6. Winter (December–March) time series of the North Atlantic Oscillation (mb), based on the normalized difference in sea level pressure between Lisbon, Portugal and Stykkisholmur, Iceland (L_n-S_n). Results are updated from Hurrell (1995, 1996).

in winter during 1955–1970 and returned northward after 1980. Hurrell argues that nearly all of the winter warming across Europe and Eurasia and cooling over the northwest North Atlantic since the mid 1970s is associated with the NAO. However, the most recent data (Figure 6) show that the winter NAO index was strongly negative for winter 1995/1996 and near neutral for winter 1996/1997 and 1997/1998. As discussed by Wallace et al. (1996) and is evident from inspection of Figure 1, the temperature signals associated with both the NAO and ENSO load positively on the COWL pattern and hence in part account for its existence. Note that in contrast to the post 1970s era, the pronounced increase in northern high latitude temperatures from 1920–1940 (Figure 3) occurred in conjunction with a sharp downward tendency in the NAO index.

Walsh et al. (1996) examine changes in sea level pressure (SLP) over the Arctic Ocean during the ‘buoy era’ 1979–1994. Their analysis shows reductions in SLP over the period 1987–1994, compared with the previous eight-year period, which are largest near the pole and statistically significant for autumn and winter. With respect to the 16-year mean, annual pressures have been below normal for every year since 1988. The difference map of mean cold season (October–March) SLP north of 30° N between the decades 1983/1984–1992/1993 and 1973/1974–1982/1983 (Figure 7) from the subsequent study of Serreze et al. (1997) clearly shows the large pressure reductions over the central Arctic Ocean (locally exceeding 4.0 mb). These changes are only part of a larger-scale shift in SLP. Pressure increases are

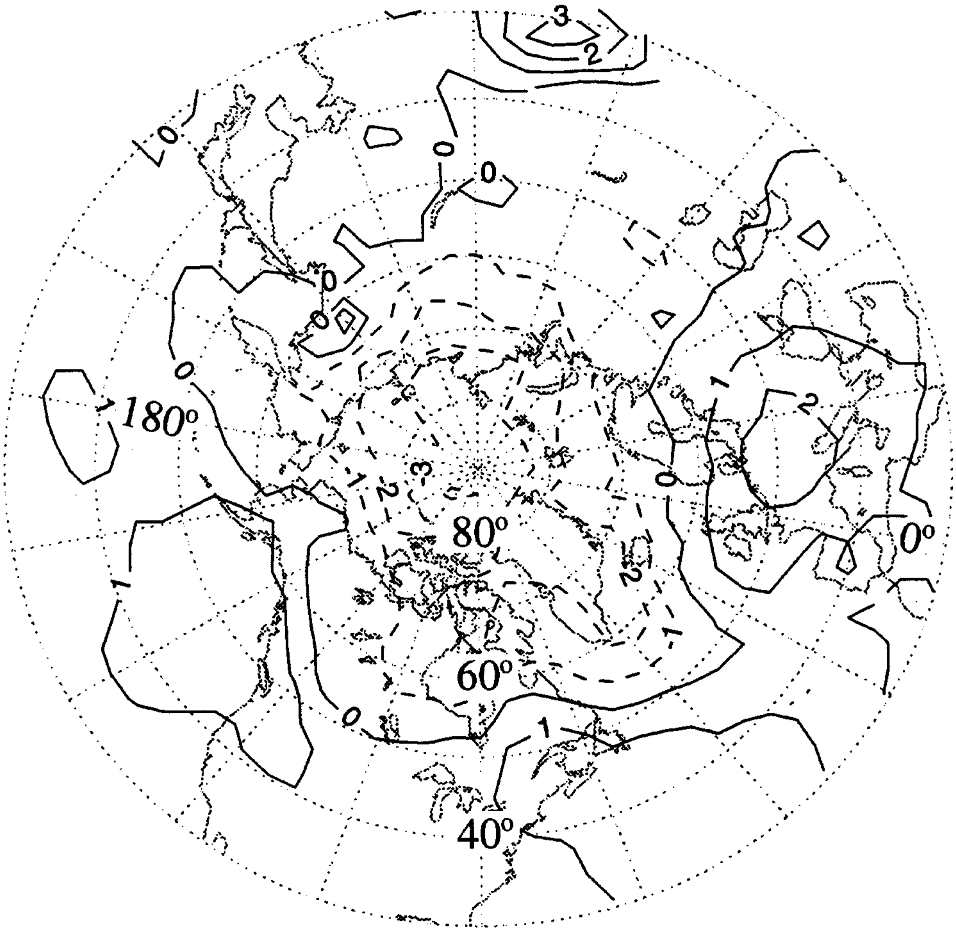

Figure 7. Difference field of mean SLP (mb) for the cold season (October–March), 1983/84–1992/93 minus 1973/74–1982/83. Positive differences are shown by solid contours with negative differences shown by dashed contours (from Serreze et al., 1997).

observed over central Europe, the northeastern Pacific, and south-central Asia over the Himalayas (in the last case these cover a small area with tight gradients and are possibly spurious, related to reduction of pressure to sea level). It is apparent that the reductions in pressure near the Icelandic Low (–2.5 mb) and higher pressures in the area of the Azores High (+1.5 mb), although consistent with the more positive NAO during the later period, are smaller than the central Arctic changes.

Thompson and Wallace (1998, 2000) find that the first Empirical Orthogonal Function (EOF) of monthly-mean SLP for November through April north of 20° N depicts a strong center over the central Arctic Ocean and weaker centers of opposing sign over the Atlantic and Pacific basins. This circulation mode, which has come to be known as the Arctic Oscillation (AO), has been generally positive

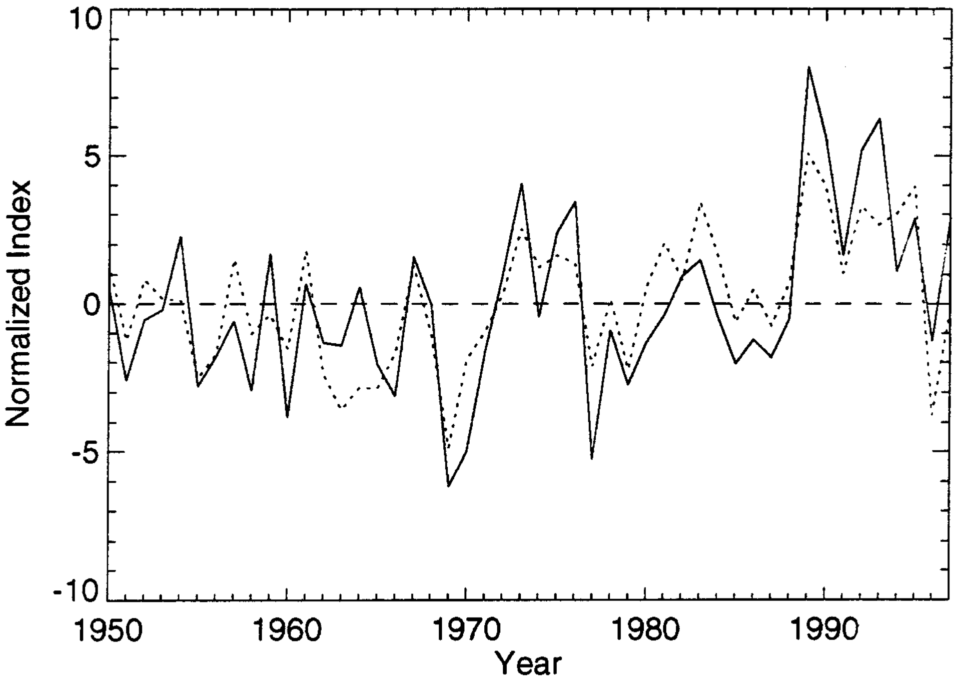

Figure 8. Time series of the Arctic Oscillation index averaged for December–March for 1950–1997 (solid line) and corresponding NAO index (dotted line), both in normalized units (AO index provided by D. Thompson, NAO index provided by J. Hurrell).

since the early 1970s. Existing evidence indicates that the NAO can be considered as a major component of the AO. For winter (December–March) averages over 1950/1951 to 1996/1997, the time series of the two indices (Figure 8) correlate at 0.79. For the full period of shared record (1899/1900–1996/1997) the correlation is 0.81, perhaps surprising given the low quality of Arctic sea level pressure data in early years.

The positive polarity of the AO is characterized by negative SLP anomalies over the Arctic which occur in association with higher temperatures over northern Siberia, and lower temperatures over the Labrador Sea, east of Hudson Bay and southern Greenland, quite similar to the pattern in Figure 1. The SLP changes shown in Figure 7 are consistent with the change in the AO. The AO pattern is found to extend from the surface to the stratosphere, hence reflecting an increase in the strength of the polar vortex. Thompson and Wallace show that the AO is more strongly correlated with Eurasia temperature than the NAO. Although the AO is a cold-season pattern, recent high latitude pressure reductions similar to those seen in Figure 7 are also observed for the warm season (April to September) (Serreze et al., 1997).

The circulation regime in middle and high latitudes has been documented by two Russian 'schools' for the entire twentieth century. The Vangengeim–Girs circulation classification (Elementary Synoptic Processes) for latitudes between about 35° N and 80° N was developed at the Arctic and Antarctic Research Institute (Girs, 1981) while that of Dzerdzeevskii (1963), termed Elementary Circulation Mechanisms, was developed at the Institute of Geography, Academy of Sciences. There is an extensive body of literature by both groups, summarized by Barry and Perry (1973) up to the early 1970s, but there has been no published intercomparison of the two approaches to our knowledge.

Based on the Vangengeim–Girs classification, Dmitriev's (1994) analysis of the frequency of cyclonic versus anticyclonic patterns over the Arctic Ocean for 1950–1992 finds that cyclonic patterns became more frequent on an annual basis after 1980. Anticyclonic patterns extending from the North Pole into Keewatin, which are climatologically most frequent in spring, decreased sharply in the 1980s in winter and summer and patterns with cyclonic circulation over the Eurasian Arctic Ocean centered near 80° N, 160° E (most common during July–October) became increasingly common in both winter and summer in the 1980s. Using an index of the zonality of the circulation between 60° N to 80° N, Dmitriev also demonstrates low zonality during 1949–1952 and 1957–1960, stronger zonal tendencies in the 1960s and mid 1970s, and persistent zonal circulation in all sectors of the Arctic for 1986–1993. This trend is most marked in the winter season. Based on data from 1900–1979, the Dzerdzeevski classification shows an annual ratio of zonal to meridional patterns of about 0.5 around 1915, 1.1 in the 1930s–1940s, 0.75 around 1960 and recovering to 0.9–1.0 in the 1970s (Savina, 1987). Changes in circulation patterns documented in these Russian studies hence appear to be broadly consistent with those in SLP and the AO index.

Serreze et al. (1997) examine changes in Northern Hemisphere extratropical cyclone activity. Based on results using an automated system detection/tracking algorithm applied to twice-daily SLP data, cyclone activity has increased north of 60° N since the mid 1960s for both the cold and warm seasons, with attendant reductions in storminess for the zonal band 30–60° N. The high latitude increases for both seasons coincide with the area of largest reductions in SLP. Time series of cold-season cyclone counts for the region north of 60° N and for 30–60° N indicate the increase in northern high-latitudes as most pronounced from the early 1980s onwards, roughly coinciding with the more pronounced upward trend in mean Northern Hemisphere temperatures, and generally positive NAO and AO index values. Counts of warm season cyclones north of 60° N also display a more general increase since 1966. Cold season activity for lower latitudes decreased from 1966/1967 to 1975/1976, but with generally high values in the subsequent decade, with low values from 1988/1989 onwards.

The spring/summer Arctic Ocean warming noted by Martin et al. (1997) and evident in the POLES data set can be considered broadly consistent with increased warm advection in this region associated with increased cyclone activity. In a similar vein, Rogers and Mosley-Thompson (1995) report that recent mild winters over north-central Asia reflect stronger westerly flow and stronger intrusions of cyclone warm sectors into this region.

The cyclone detection/tracking algorithm used by Serreze et al. has been modified for application to the more consistent 6-hourly fields from the NCEP/NCAR reanalysis. Resulting time series of cyclone counts and intensity north of 60° N for standard calendar seasons over the period 1958–1997 are provided in Figure 9. Intensity is based on the Laplacian of SLP at the cyclone centers, which provides a better measure of intensity than does central pressure. Both variables exhibit large interannual variability, superimposed on upward trends. With the exception of autumn cyclone counts, all trends are statistically significant to at least the 95% confidence level. When data are examined for higher latitudes (north of 70° N and north of 80° N), the trends in counts become significant only for summer, but those for intensity remain significant for all seasons (except for north of 80° N in autumn). While questions remain regarding the extent to which these results reflect remaining analysis biases due to changes in the amount and quality of assimilation data, the results seem to reconfirm that overall, Arctic cyclones are becoming both more common and more intense, implying increased poleward heat transport. The summer increases, however, are not clearly related to the NAO or AO, which are largely cold season phenomena.

2.3. SNOW COVER

Weekly National Oceanic and Atmospheric Administration (NOAA) Northern Hemisphere charts of snow covered area (SCA), derived primarily from analysis of visible-band satellite imagery, are available since 1972 (Robinson et al., 1993, 1995). Based on these data, the areal extent of Northern Hemisphere snow cover ranges from a winter maximum of about 46×10^6 km^2 to a summer minimum of 4.0×10^6 km^2, but with large interannual variability. Mean annual SCA is 25.3×10^6 km^2, unevenly divided between Eurasia (14.7×10^6 km^2) and North America (10.6×10^6 km^2). The majority of snow-covered lands lie north of 50° N. With its large extent and seasonal amplitude, high albedo (upwards of 0.80 when fresh and unforested) and low thermal conductivity, snow cover is a key element of the climate system. Through the temperature-albedo feedback mechanism, changes in snow cover, along with sea ice extent, are expected to contribute to polar amplification of externally-driven climate warming.

NOAA data analyzed through August 1998, presented as monthly anomalies and twelve month running means of Northern Hemisphere SCA (Figure 10a), show generally (but by no means always) above-average coverage from the beginning of the record through the mid 1980s. Within this period, snow cover was particularly extensive in the 1970s and mid 1980s. By comparison, the late 1980s through August 1998 has been a period of generally subnormal SCA. This pattern is seen over both North America (Figure 10b) and Eurasia (Figure 10c) although there appears to have been a partial recovery during the mid 1990s. The difference in annual

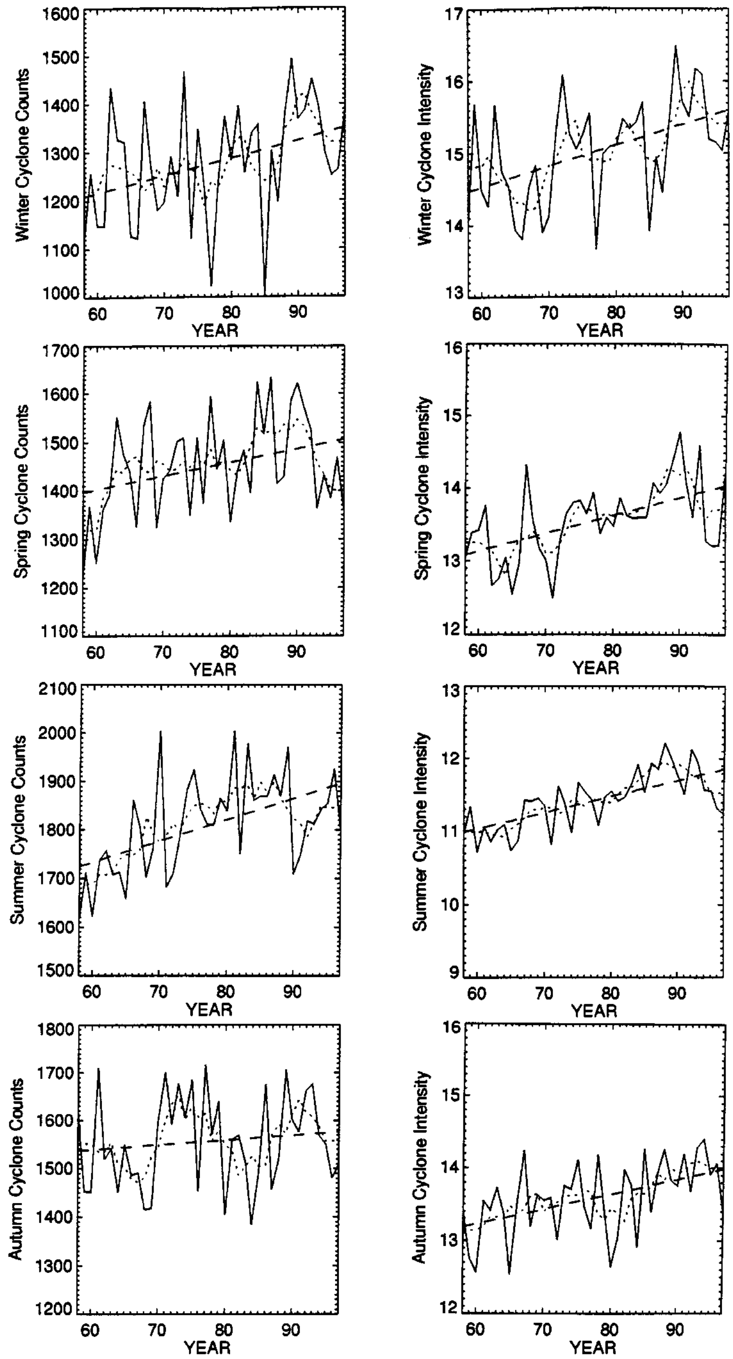

Figure 9. Time series of cyclone counts and intensity north of 60° N by season (monthly values, five-year running means and least-squares fit), 1958–1997, based on NCEP/NCAR reanalysis fields. Intensity is based on the Laplacian of sea level pressure at the cyclone centers and has units of 10^5 mb km^{-2}.

(a)

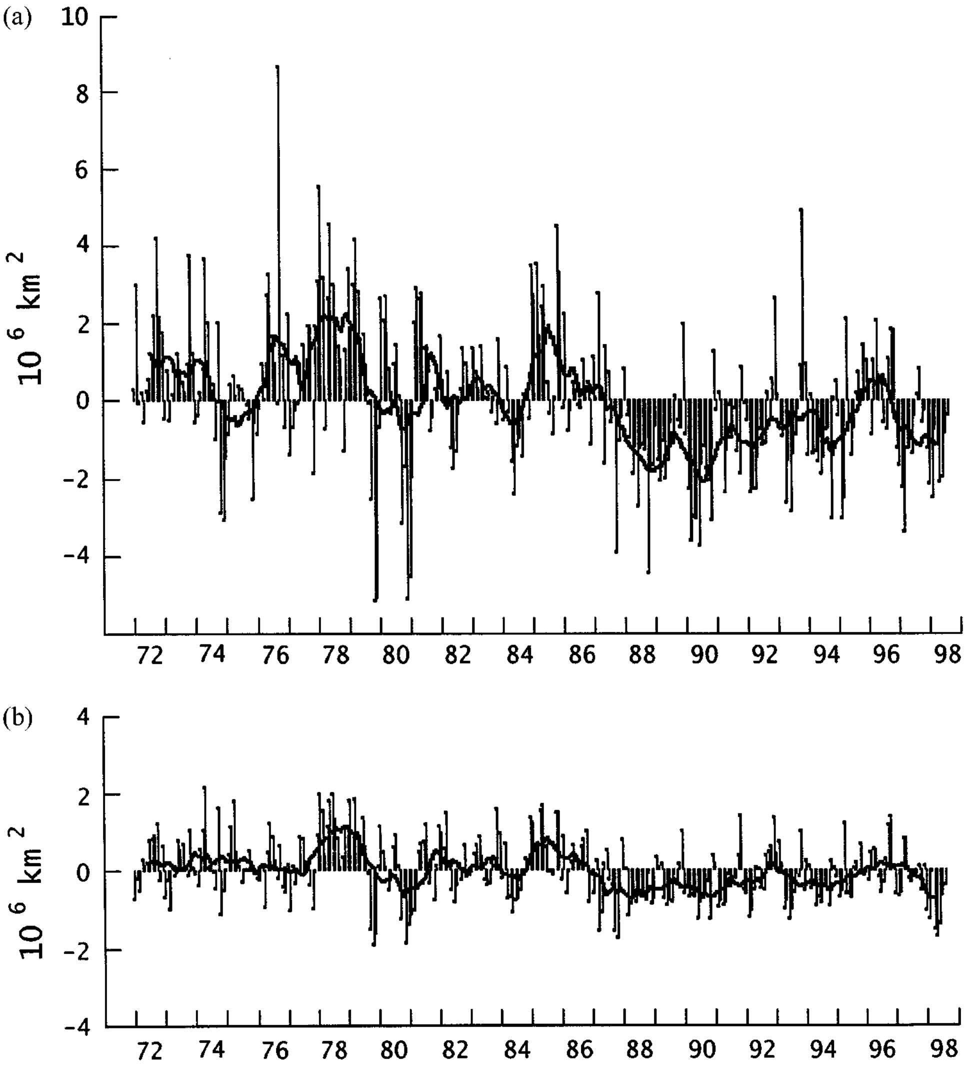

Figure 10a,b.

Figure 10. Monthly anomalies and 12-month running anomalies of snow cover extent (106 km^2) over: (a) Northern Hemisphere lands; (b) North America; (c) Eurasia. Results for the Northern Hemisphere and North America include Greenland. The data record extends from January 1972 through August 1998 (updated from Robinson et al., 1993).

means between 1987–present and the preceding period is statistically significant and the largest changes have occurred during spring and summer. Overall, Northern Hemisphere annual SCA has declined by about 10% since 1972 (Groisman et al., 1994b).

(c)

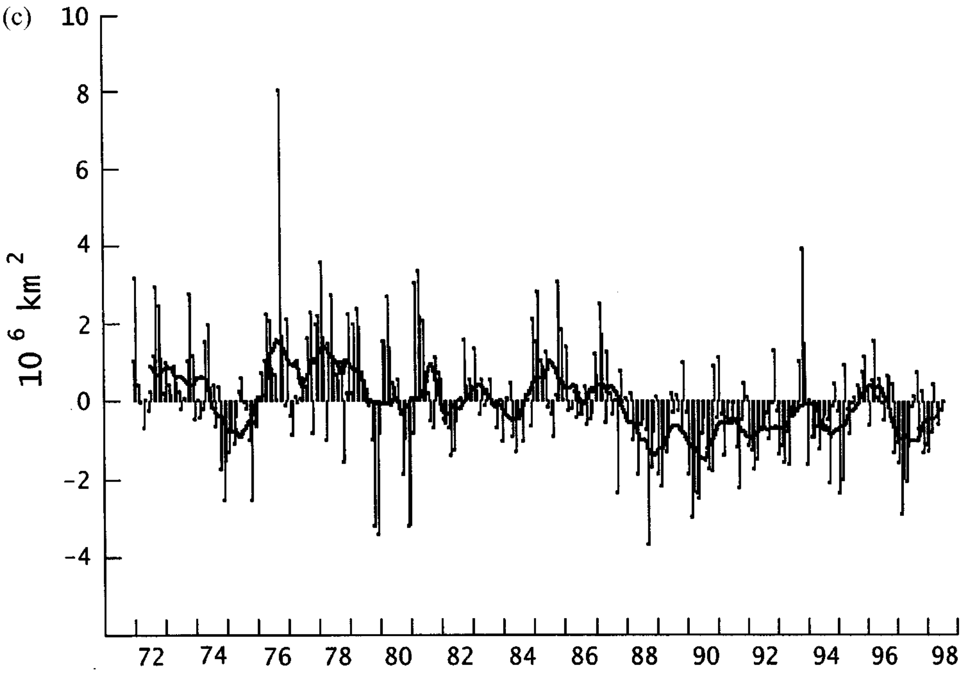

Figure 10c.

Longer regional time series are available based on station records and reconstructions. Using station data, Brown and Braaten (1998) show that for the period 1946–1995, snow depths during January–March decreased across most of Canada, with the largest decreases in March. Decreases are most prominent in the Mackenzie Basin, Prairies, and lower St. Lawrence Valley. Increases were noted only on the east coast. Snow cover duration also declined over most of western Canada and in the Arctic in summer. The changes are characterized by a sharp transition to lower snow depths in the mid 1970s.

Brown and Goodison (1996) examine SCA over Canada from 1915–1992 for four regions (West Coast, Western Prairie, Southern Ontario/Quebec and the Maritimes). Results are based on reconstructions employing station records of snow depth, snowfall and maximum temperature along with a simple mass balance model whereby snowmelt is estimated via a calibrated temperature index. The model was calibrated using observed data for the period 1955–1992. No statistically significant long-term trend in SCA was found in any of the four regions, but the data suggest that winter snow cover increased and spring snow cover decreased over much of southern Canada during 1955–1992. A notable regional feature was a systematic reduction in snow cover over the Canadian Prairies since about 1970 in winter and spring.

Time series for their four regions were used along with other regional time series and NOAA satellite records in a stepwise regression to estimate North American

SCA back to 1915. The reconstructions suggest that North American winter SCA has exhibited a gradual increase of $11.0x10^3$ km^2 yr^{-1} while spring snow cover has decreased by about 6×10^3 km^2 yr^{-1}. These represent a change of $<10\%$ of the current mean SCA.

Meshcherskaya et al. (1995) analyze trends in winter snow depth for 1891 to 1992 in the agricultural regions of the former Soviet Union. There are decreases in European Russia, except around Sverdlovsk and Ufa. No details are given on recent trends. Fallot et al. (1997) examine average winter season (November–April) snow depths at 283 stations across the FSU up to 1983/1984, and report a tendency for depths to have increased in European Russia north of 63° N from 1945–1950 to the early 1980s and in western Siberia after about 1960 or 1970.

In summary, satellite records indicate that since 1972, Northern Hemisphere annual snow cover has decreased by about 10%, largely due to spring and summer deficits since the mid 1980s over both continents. In turn, there is evidence that for Canada, there has been a general decrease in snow depth since 1946, especially during spring, and that winter depths have declined over European Russia since the turn of the century. However, reconstructions for Canada suggest that while there has been a general decrease in spring SCA since 1915, winter SCA has increased. Winter snow depths over parts of Russia also appear to have increased in recent decades. The common thread between studies that have examined seasonality is an overall reduction in spring snow cover.

With regard to the decrease of Northern Hemisphere annual SCA by approximately 10% since 1972, Vinnikov and Robock (1998) find that trends of such magnitude are rare events in a 1000-year simulation by the GFDL global climate model. Groisman et al. (1994a,b) also argue that the reduction in Northern Hemisphere spring snow cover can be related to an increase in snow cover radiative feedback, accounting for part of the observed increases in spring temperatures.

2.4. PRECIPITATION AND P-E

Models predict that the enhanced temperature response of the Arctic to anthropogenic greenhouse forcing will be attended by increases in precipitation during winter, related to higher atmospheric water vapor content (precipitable water) and poleward vapor transport (Kattenberg et al., 1996). Changes in the high-latitude terrestrial hydrologic budget, including the amount and seasonality of precipitation, evaporation, snow water equivalent, the timing of snow melt and runoff may influence terrestrial ecosystems. As discussed in Section 3, river runoff into the Arctic Ocean as well as the balance between precipitation and evaporation (P-E) over the Arctic Ocean itself may impact the sea ice cover, freshwater transports into the North Atlantic and deep ocean convection.

Assessing changes in the atmospheric components of the northern high latitude hydrologic budget is difficult, even for 'base' variables such as precipitation. The station network is fairly sparse and there are significant problems of undercatch of solid precipitation (Woo et al., 1983), although some investigators (e.g., Groisman et al., 1991; Groisman and Easterling, 1994) have attempted to correct for gauge biases. Based on available data, annual precipitation as evaluated for the period 1900–1994 increased over both North America and Eurasia (Nicholls et al., 1996). For North America, positive trends in annual precipitation as well as snowfall are most apparent (up to a 20% increase) during the past 40 years over Canada north of 55 °N (Groisman and Easterling, 1994). For the former Soviet Union, most of the increases occurred during the earlier part of the 20th century and are larger during winter than for summer, with a tendency for reduced precipitation in some areas since the middle of the century (Groisman et al., 1991). As summarized for zonal bands, annual precipitation for the period 1990–1995 increased for the region 55° N–85° N, with the largest changes during autumn and winter (Figure 11). The recent analysis of Dai et al. (1997) confirms these results. The extent to which reported recent increases in Siberian river discharge (Semiletov et al., 1999) relate to these precipitation trends remains to be fully resolved.

This apparent correspondence with models in terms of the observed increases must be interpreted in the context of sampling and undercatch biases in the observational record, potential impacts of natural variability in atmospheric circulation, as well as apparent biases in model simulations of present-day precipitation. Comparison of approximately two dozen climate model simulations from the Atmospheric Model Intercomparison Project (AMIP) show a definite tendency for the models to overestimate Arctic precipitation (Walsh et al., 1998a). However, a subset of the models (e.g., Colorado State University (CSU), Japan Meteorological Agency (JMA), Main Geophysical Observatory (MGO), University of California Los Angeles (UCLA), University Global Atmospheric Modelling Project (UGAMP)) simulate Arctic precipitation amounts that agree closely (within about 10%) with observational estimates. Most of these models include the evaporation of falling rain and variable soil water capacities among their suites of parameterizations.

From a hydrologic viewpoint, P-E is arguably more important than precipitation by itself. Two studies (Walsh et al., 1994; Serreze et al., 1995a) have utilized the network of northern high latitude rawinsonde stations to examine P-E averaged over the Arctic Basin north of 70° N via the 'aerological' approach. Based on data from the early 1970s through the early 1990s there are no obvious trends, with a mean annual value around 16–17 cm. Recent updates through the middle of 1996 also reveal no trends. Parallel efforts using analyzed wind and moisture fields from the NCEP/NCAR and European Center for Medium Range Weather Forecasts (ECMWF) reanalysis archives yield somewhat higher mean annual values (18–19 cm) (Cullather et al., 2000), but also no trends (Bromwich, personal communication, November 1997).

The major transport pathway of water vapor into the Arctic Basin is near the prime meridian (10° W–50° E) in association with the North Atlantic cyclone track. The winter season (DJF) meridional moisture transports both along this sector and averaged over all longitudes exhibit a positive relationship with the phase of the

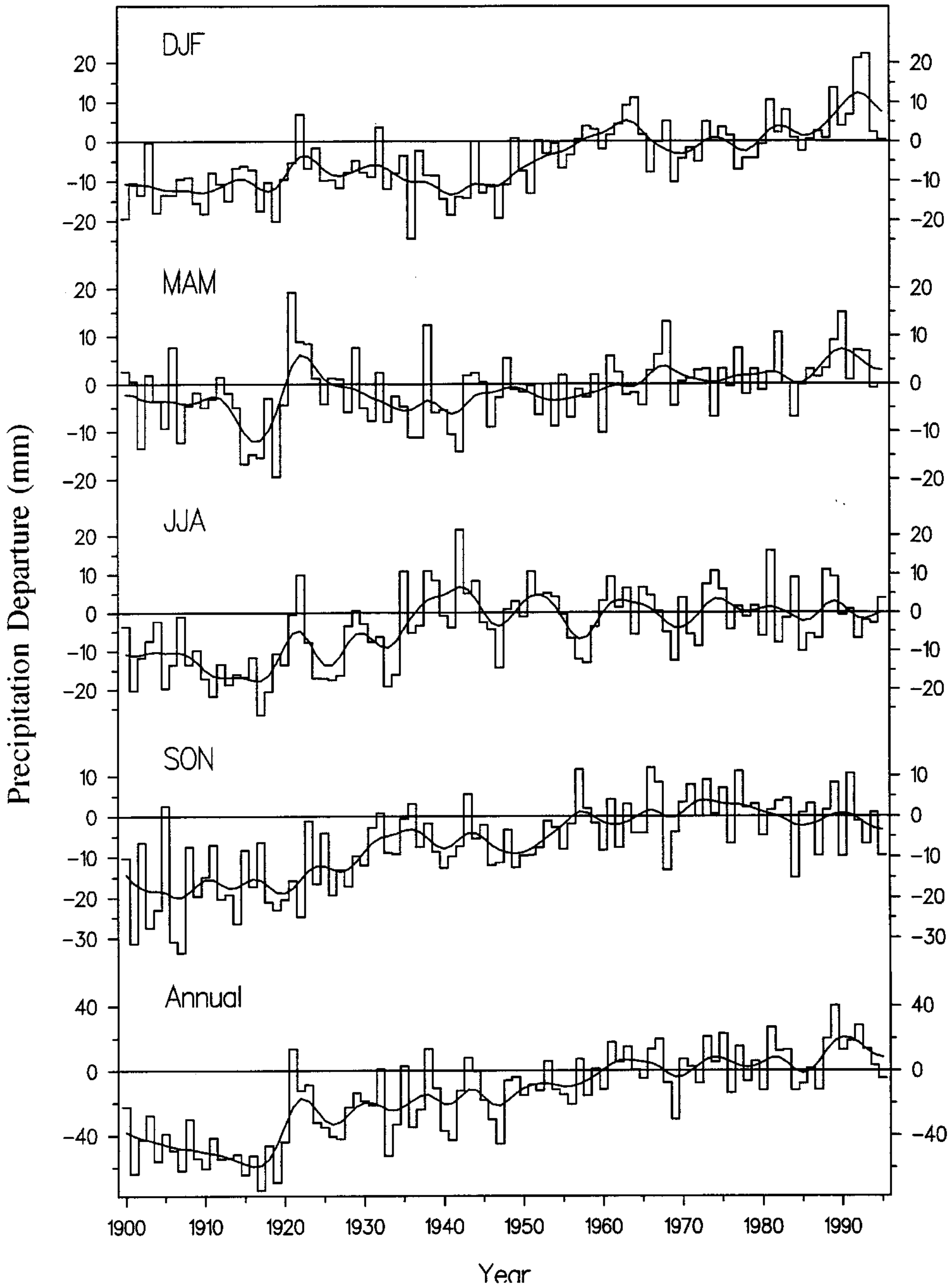

Figure 11. Annual and seasonal precipitation anomalies (55–85° N) for 1900–1995 evaluated with respect to 1951–1980 means (mm). The smoothed line represents results from a nine-point low-pass filter. Results are based on updates to the Eischeid et al. (1995) data set.

NAO (Dickson et al., 2000). As noted earlier, the NAO has been largely positive in recent decades. Nevertheless, winter P-E shows no trend, apparently reflecting the large variability in circulation.

3. Ocean

3.1. SEA ICE

The Arctic sea ice cover ranges in areal extent from a maximum of about 14.8×10^6 km^2 in March to a minimum of about 7.8×10^6 km^2 in September (Parkinson et al., 1987). The thickness is highly variable with area mean values of 3–5 m in the central Arctic. During winter, the ice cover inhibits heat exchange between the cold atmosphere (down to –40 °C) and the relatively warm surface of the Arctic Ocean. Along with its large areal coverage, sea ice has a high albedo of up to 80% during spring when covered with fresh snow but still typically 50% during summer melt (Robinson et al., 1992), strongly limiting absorption of solar radiation. These effects help to maintain the Arctic as a global heat sink. Sea ice (other than young first year ice) is essentially fresh water, with typical salinities of 1–6 ppt. Changes in the ice flux out of the Arctic, primarily via Fram Strait, may influence the oceanic convective regime and deepwater formation in the Greenland Sea–North Atlantic (Rudels, 1989), affecting the global thermohaline circulation.

Along with low air temperatures, the sea ice owes its existence to a low-salinity layer (as low as 29 ppt at the surface) extending to about 200 m depth, maintained by river runoff, inflow of low-salinity waters (31–33 ppt) through Bering Strait, and a P-E excess over the Arctic Ocean itself (Aagaard and Carmack, 1989). The surface layer is underlain at 200–900 m in depth by the Atlantic layer, derived from inflow into the Norwegian Sea from the North Atlantic Current. This layer is relatively warm with temperatures above 0 °C, and, if brought to the surface would quickly melt the ice cover. However, at the low water temperatures of the Arctic Ocean, the vertical density structure is determined by salinity, rather than temperature. This limits the depth of vertical mixing of seawater to typically 40–70 m, allowing sea ice to form readily in winter and inhibiting summer melt.

Based on analysis of the satellite passive microwave record from the Nimbus-7 Multichannel Microwave Radiometer (SMMR) through 1987, Gloersen and Campbell (1991) demonstrate a small but significant downward trend in Arctic sea ice extent. Chapman and Walsh (1993), using a longer record (1961–1990) based on weekly U.S. Navy/NOAA National Ice Center charts since 1973 and regional sea ice data sources for earlier years, confirm a downward trend. Johannessen et al. (1995) subsequently found that this downward trend has increased since about 1989. This view is reinforced by more recent work (Bjorgo et al., 1997), in part driven by concerns over errors in sea ice retrievals from passive microwave data and problems in blending the earlier SMMR records (1978–1987) (Parkinson and

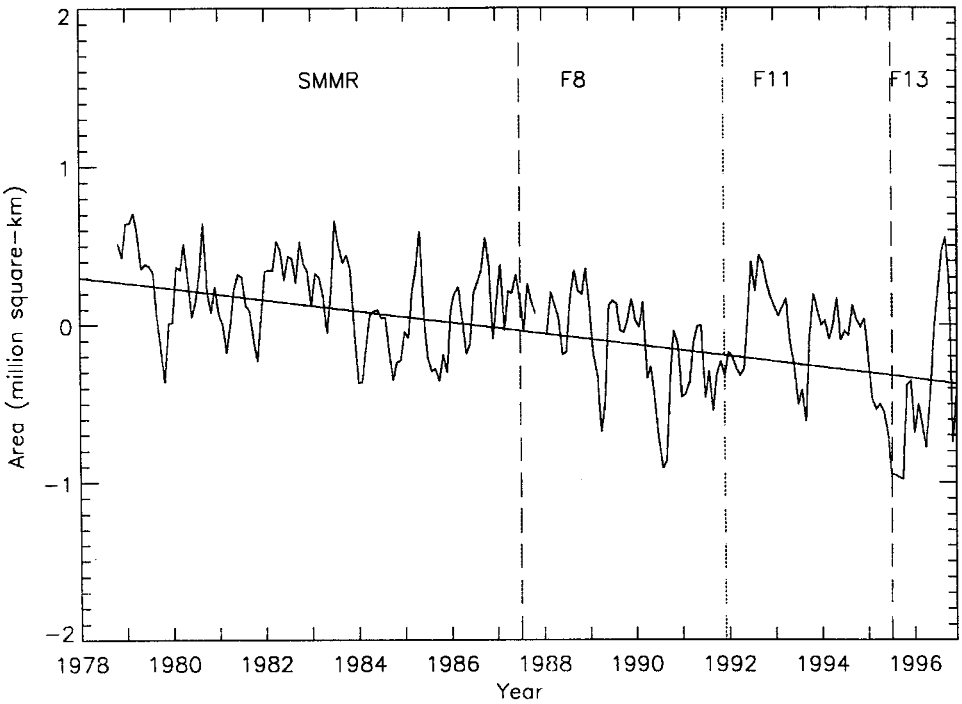

Figure 12. Arctic sea ice extent anomalies (monthly means and least squares fit), 1979–1996 (10^6 km^2). The vertical dashed lines mark the period of coverage by SMMR and the more recent SSM/I instrument aboard the F8, F11 and F13 platforms. Data represent retrievals using the NASA Team Algorithm (cf. Cavalieri et al., 1997). Data were provided by NASA and processed at the National Snow and Ice Data Center, Boulder, CO.

Cavalieri, 1989) with the more recent time series (1987 onwards) from the Defense Meteorological Satellite Program Special Sensor Microwave/Imager (SSM/I). The most recent study using passive microwave data through 1996 (Cavalieri et al., 1997) shows Arctic sea ice extent decreasing by 2.9 +/- 0.4% per decade but Antarctic sea ice extent increasing by 1.3 +/- 0.2% per decade. Also based on the passive microwave time series, Smith (1998) shows that these ice reductions have been accompanied by a general increase in the length of the ice melt season.

At least two studies (Serreze et al., 1995b; Maslanik et al., 1996) have examined the seasonality and forcing mechanisms of recent changes in ice extent. As seen in the time series of Northern Hemisphere ice extent in Figure 12, ice extent exhibits large variability, superimposed on an overall downward trend. Further inspection of the time series shows that the annual trend is strongly driven by trends in late summer and early autumn (Figure 13). Extreme minima, unprecedented within the passive microwave record, are found during 1990 and 1995. These reflect primarily reduced ice cover over the Laptev and East Siberian seas where (based on data through 1995) ice extent has decreased fairly steadily since about 1990.

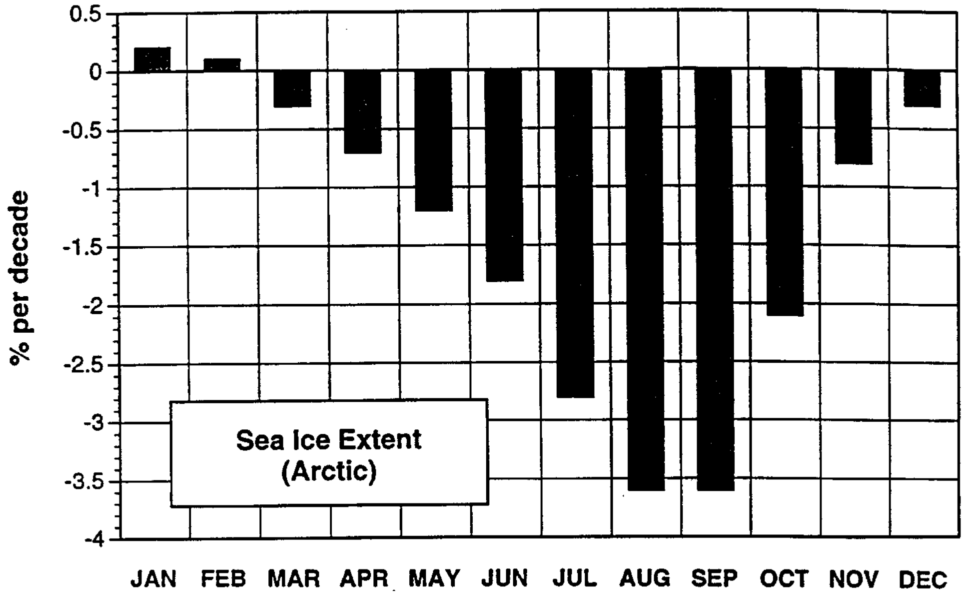

Figure 13. Monthly trends in Arctic sea ice extent (1979–1995) in % per decade.

As reviewed in Section 2.2, extratropical cyclone activity for both the warm and cold half of the year has increased over the central Arctic Ocean, with associated reductions in sea level pressure (Figure 7), attended by higher spring and summer temperatures over the Arctic Ocean (Figure 4). Comparison of the trends in Figure 4 and Figure 13 suggests that the overall pattern of enhanced late summer/early autumn sea ice reductions relate in part to these antecedent temperature anomalies promoting earlier melt onset. This is consistent with the view of Maslanik et al. (1996) that the location of increased cyclone activity over the central Arctic Ocean advects warm southerly winds into the Laptev and East Siberian Seas. However, as Maslanik et al. also point out, southerly winds also transport ice away from the coast. The relative importance of these contributions remains to be quantified.

However, Figure 12 shows that total ice extent was well above average in 1996. The most recent passive microwave data show record low ice extents ice in the Beaufort Sea in summer 1998. These results are consistent with reports from the manned camp of the Surface Heat Budget of the Arctic (SHEBA) experiment of extensive melt, thin ice, low surface salinities and an anomalous northward ice limit. However, in terms of total ice extent, this anomaly is at least partly offset by more extensive ice on the Eurasian side of the Arctic, hence contrary to the general pattern seen in the 1990s (Maslanik et al., 1999)

With regard to longer-term (century-scale) changes, Zakharov (1997) has shown a substantial decrease of sea ice coverage in the eastern North Atlantic during the twentieth century. This trend is also apparent in the charts of the Danish Meteorological Institute (Walsh et al., 1999), although the data used in these syntheses are primarily for the spring-summer portion of the year. Vinje and Colony (1999) extend the time series back several centuries in the vicinity of the Norwegian Sea. Large decreases of sea ice extent since the 1890s are apparent. The updated Koch Index (Koch, 1945) of sea ice near Iceland also indicates that the twentieth century has been relatively ice-free near Iceland in comparison with the previous century. Because these results are based primarily on reports from ships and land stations, their compatibility with more recent satellite-derived datasets is open to question. However, the recent century-scale decrease of sea ice in the North Atlantic is a common theme in the results, implying that the decrease is most likely real.

3.2. OCEAN CIRCULATION

Results from several recent oceanographic cruises indicate that the influence of the Atlantic Water at 200–900 m depth (Section 3.1) has become increasingly widespread and intense. Data collected during the cruises of the U.S.S. Pargo and Henry Larson in 1993 (Carmack et al., 1998; McLaughlin et al., 1996; Morison et al., 1998) and the summer 1994 Arctic Ocean Section of the Polar Sea and Louis S. St. Laurent (Carmack et al., 1998) all indicate that the boundary between the eastern and western Arctic Ocean halocline types now lies roughly parallel to the Alpha and Mendeleyev Ridges (AMR). This suggests that the areas occupied by the Atlantic water types have increased by up to 20%. Figure 14 summarizes salinity and temperature measurements from the 1993 cruise track of the U.S.S. Pargo. The Pargo data and Gorshkov climatology agree in the Canada Basin. However, the salinity in the Makarov Basin and over parts of the AMR is substantially higher in the recent measurements, indicating an influx or more saline Atlantic-derived water into the Makarov Basin. The sharpness of the front between the Atlantic-derived and Pacific-derived waters is captured by the sail CTD (conductivity-temperature-density) data.

The greater Atlantic influence is also manifest in warm cores observed over the Lomonosov and Mendeleyev ridges, with temperatures over the Lomonosov Ridge greater than 1.5 °C. This is also illustrated in Figure 14 where the temperature at 100 m is substantially higher over the Lomonosov Ridge at 2100 and 2800 km into the track. The maximum warming actually occurs between 150 m and 200 m depth. This is above the center of the observed warm core at 250 m because the depth of the temperature maximum is also less than in the climatology. Carmack et al. (1998) and McLaughlin et al. (1996) also observed an Atlantic Layer temperature increase over the Mendeleyev Ridge. Results from the Transarctic Acoustic Propagation (TAP) experiment conducted in April 1994 also suggest warmer waters in the Atlantic layer (Mikhalevsky et al., 1995). Historical data give no indication of such warm cores and show a temperature over the Lomonosov Ridge nearly 1 °C lower. The recently-prepared digital atlas of Russian and U.S. hydrographic data (Environmental Working Group, 1998) confirms that no temperatures greater than 1 °C were observed during numerous investigations between

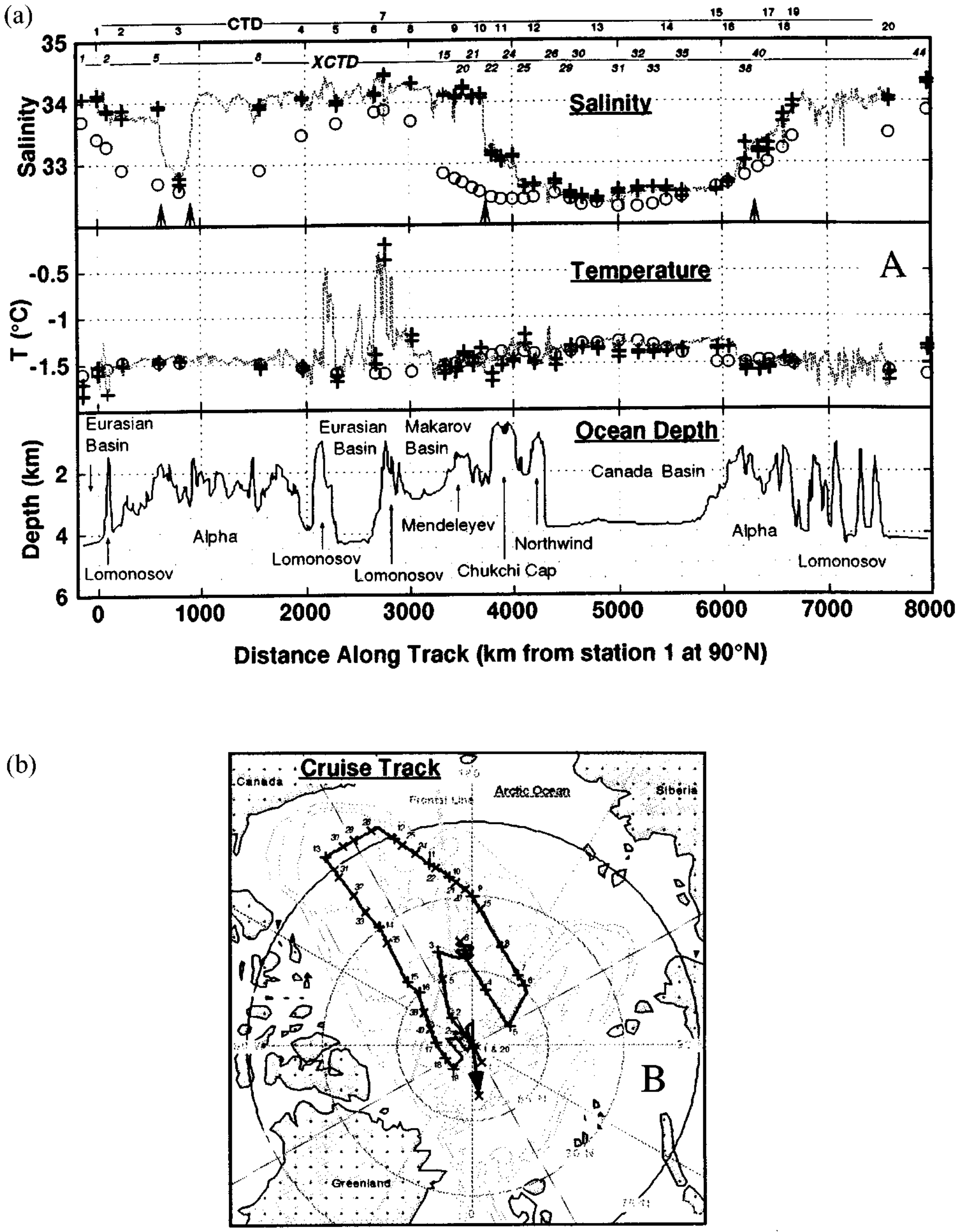

Figure 14. (a) Cruise track, salinity and temperature from CTD and SSXCTD data at 100 and 104 m (+) and sail CTD at 103 m (light lines) from the 1993 cruise of the U.S.S. Pargo. Salinity and temperature are interpolated to the track from the Gorshkov 100 m climatology (○). Ocean depth and key oceanographic regions are shown in the lower panel. Points corresponding to frontal line intersections are indicated on the salinity plot by the upward-pointing arrows. (b) Location map showing cruise track. The CTD and SSXCTD station numbers correspond to those in Figure 14a (based on Morison et al., 1998).

1950 and 1989. This warming may have started in the late 1980s or early 1990s. Cruise data from the Oden in 1991 (Anderson et al., 1994; Rudels et al., 1994) show slight warming of the upper ocean near the pole and Quadfasel (1991) reports higher-than-usual temperatures in the Atlantic water inflow in 1990.

The extent to which these Atlantic layer changes may reflect atmospheric and wind forcing is being investigated. Swift et al. (1998) suggest that recent warm winters in the Norwegian Sea associated with the generally positive phase of the NAO have led to increased advection of warmer waters into the Arctic (see also Morison et al., 1998). It is also important to note that other oceanic changes have been observed. Steele and Boyd (1998) show a retreat of the Cold Halocline Layer (CHL) from the Eurasian Basin into the Makarov basin, which they have attributed to the effects of anomalous wind forcing shifting the 'injection point' of fresh riverine shelf waters. Steele and Boyd (1998) also summarize changes in Pacific-influenced water types. As a cautionary note, Grotefendt et al. (1998) argue that while the warming of the Atlantic Layer as compared to Russian climatologies is significant, about half of it can be attributed to different methods by which the earlier and later data sets were obtained.

4. Glaciers and Permafrost

4.1. GLACIER MASS BALANCE

As reviewed by Warrick et al. (1996), global mean sea level has risen by 10–25 cm over the last 100 years. Although consistent with thermal expansion related to higher global air temperatures, the sea level rise may be in part due to the melting of glaciers, ice caps and ice sheets. Observational evidence is insufficient to say with certainty whether the mass balances of the Greenland and Antarctic ice sheets have changed. Based on data from 1979–1994, Abdalati and Steffen (1997) show a tendency towards increased melt area over the Greenland ice sheet between 1979–1991, ending abruptly in 1992, possibly as a result of the stratospheric dust from Mt. Pinatubo. However, the overall global mass balance of 'small' mountain and subpolar glaciers has been negative over the past several decades (Meier, 1984; Warrick et al., 1995; Dyurgerov and Meier, 1997).

Based on a comprehensive data set enriched by addition of glaciers for the Arctic islands, small glaciers around Antarctica and the Greenland ice sheet and mountainous areas of Siberia, central Asia and the Caucasus (Dyurgerov and Meier, 1997), the area-weighted global mass balance of small glaciers evaluated for the period 1961–1990 is –130+/–33 mm, or 0.25+/–0.10 mm a^{-1} in sea level equivalent. This represents approximately 16% of the average rate of sea level rise in the past 100 years. This contribution to sea level rise has increased greatly since the middle 1980s, in broad agreement with the global air temperature record. Area-weighted balances have been positive only for the European sector. Over the period 1961–1990, the contribution of sea level rise to the melt of small glaciers is estimated at about 7.36 mm. Of this, the Arctic Islands contribute 1.36 mm (about 18% of the total). Alaska makes a smaller contribution of 0.54 mm (7%). The largest contribution has been from Asia of 3.34 mm (45%).

Figure 15 summarizes the mass balance record totaled over all small Arctic glaciers included in the Dyurgerov and Meier (1997) data set updated through 1993, with further breakdowns for Canada and Svalbard. Results are presented as annual mass balance estimates (the net gain or loss of ice over a budget year) and cumulative balances (the summation of the annual balances) in terms of water volume (km^3). A more detailed summary is provided in Table I, which gives comparisons between balances for Alaska, the Arctic Islands, Svalbard, Europe, Greenland, Asia, the Northern Hemisphere and the globe as a whole, also updated through 1993. Results are given by year and averaged over the period of record.

There is no apparent trend in annual mass balance for the Arctic average (Figure 15) but annual values have been generally negative. As such, the cumulative balance exhibits a strong downward tendency. Data averaged for Canadian glaciers reflect the Arctic-wide results. Negative annual balances have been particularly persistent for small glaciers on Svalbard. As seen in Table I, balances have also been negative for Alaska. Balances were the most negative over of the period of record for the Arctic Islands in 1991 and 1993.

It is stressed that conditions for individual glaciers vary. Dowdeswell et al. (1997) examine 40 Arctic ice caps and glaciers with records extending back to the 1940s. They find that, while most Arctic glaciers have experienced predominantly negative balances over the past few decades, some, such as in the montane parts of Scandinavia and Iceland have been positive due to increased winter precipitation.

4.2. PERMAFROST

Permafrost is perennially frozen ground which underlies 20–25% of the exposed land surface of the earth in regions with cold climates. Time scales for changes in permafrost thicknesses (and thus occurrences and distribution) range from decades to millennia (Lachenbruch et al., 1982). Permafrost is covered by the active layer (the top layer of ground, typically <1 m in thickness, that freezes and thaws annually) where biological activity occurs.

Most permafrost in high latitudes contains excess ice in the top ten meters in the form of lenses, irregular masses and wedges arranged in polygonal patterns. These form the ‘glue’ holding the permafrost together. When the excess ice thaws, the surface collapses forming thermokarst, an irregular topography of mounds, pits, troughs and depressions that may or may not be filled with water. This can result in the total destruction of ecosystems and their conversion to other types of ecosystems. Warming of the Arctic should result in higher soil and permafrost temperatures, northward movement of the permafrost boundary, a longer growing season and possibly a deeper active layer. Thawing of permafrost could accelerate

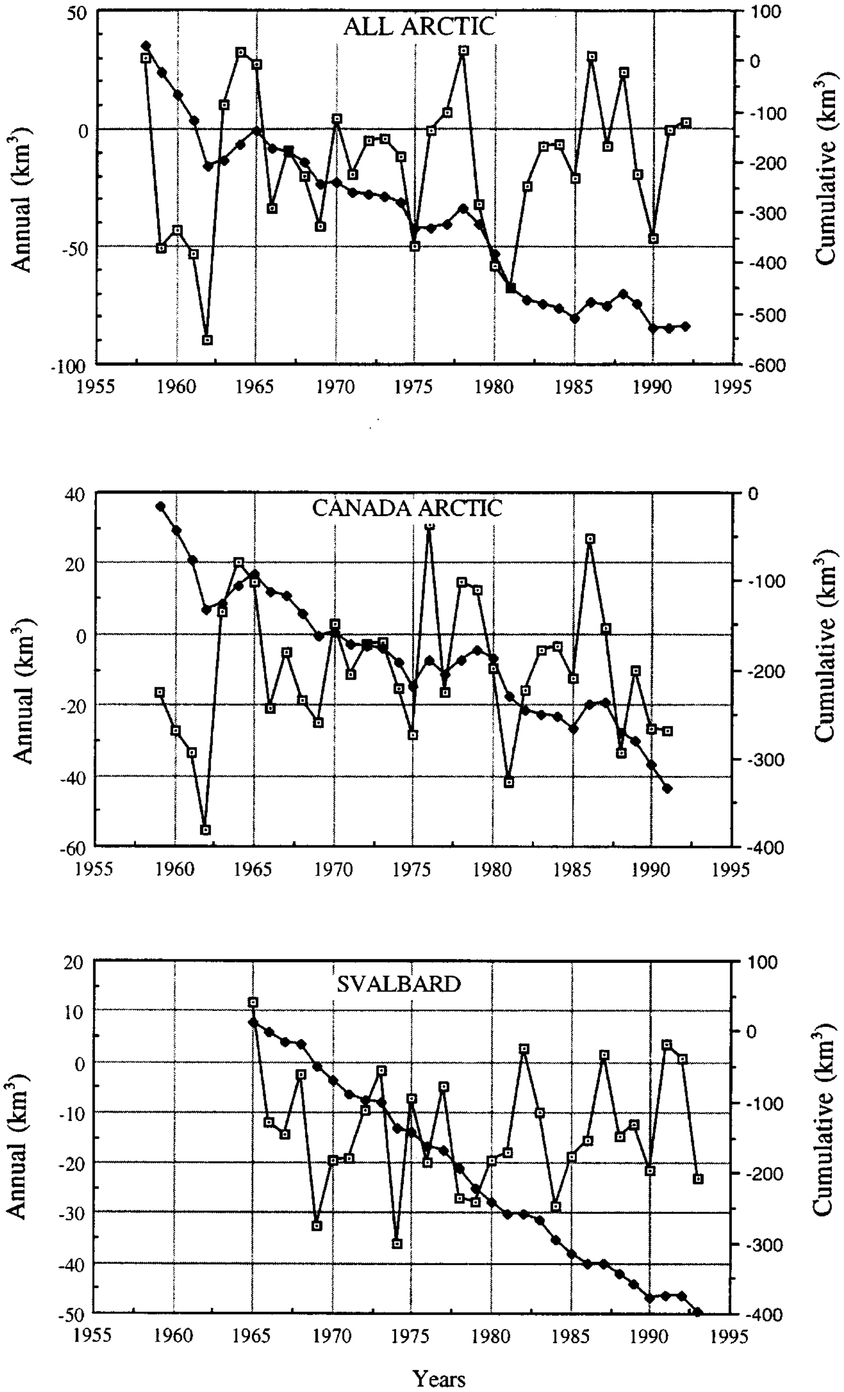

Figure 15. Area-weighted mass balance (km^3) expressed as annual (squares) and cumulative annual (diamonds) values for small glaciers. Results are for the entire Arctic, the Canadian Arctic and Svalbard (based on the Dyurgerov and Meier (1997) data set).

TABLE I

Regional averages of glacier mass balances by year (km^3) and averaged over the record length (km^3 and mm water averaged over glacier areas)

Region	Alaska	Arct. Islands	Svalbard	Europe	Green-land	Asia	Northern hemisphere	Global
Area $\times 10^3$ km^2	74.7	244.5	36.6	18	70	118.4	525.6	680
Years	B. km^3	*B. km^3	B. km^3	B. km^3	B. km^3	B. km^3	B. km^3	B. km^3
1961		–42	–42	–3	–41	–69	–106	–195
1962		–92	–42	3	–78	–79	–207	–314
1963		6	–4	–7	18	43	104	24
1964		31	–4	–8	38	23	41	173
1965	30	22	8	16	23	–34	–52	48
1966	6	–27	–14	–3	–27	–14	–48	–78
1967	–61	–11	–18	15	1	40	–42	–50
1968	–12	–33	–16	6	–35	–57	–109	–147
1969	–32	–26	–32	–10	–35	–13	–107	–140
1970	70	5	–20	–8	18	–9	18	18
1971	14	–22	–19	1	–9	–69	–90	–99
1972	–44	–6	–10	–2	33	–29	–77	–41
1973	20	–4	–2	6	–17	–89	–72	–82
1974	–80	–26	–35	5	–12	–76	–133	–165
1975	–1	–25	–7	12	–9	–65	–120	–99
1976	–61	–6	–19	–1	50	–61	–92	–33
1977	65	20	–9	11	–17	–93	–51	–53
1978	27	11	–27	12	15	–114	–83	–76
1979	–59	–61	–15	4	18	–43	–186	–192
1980	100	–81	–22	6	–7	–76	–74	–86
1981	–5	–45	–19	13	–28	–27	–111	–127
1982	14	3	1	–2	–15	–63	–157	–134
1983	19	0	–12	7	34	–49	–63	–5
1984	–29	–14	–29	9	0	–19	–64	–58
1985	39	–31	–22	–6	–19	2	–40	–96
1986	–13	33	–14	–2	53	–63	–95	–17
1987	13	2	4	7	16	–3	–43	–9
1988	29	–43	–19	–16	–48	–27	–20	–158
1989	–105	31	–18	17	–14	–24	–194	–137
1990	–112	–57	–31	–1	–33	–47	–261	–320
1991	–19	–126	–8	–14		–51	–124	–167
1992	–16	–52	–2	20		6	–79	–105
1993	–80	–154	–13	18		27	–311	–419
Years period	29 1965–93	34 1961–93	34 1961–93	34 1961–93	30 1961–90	33 1962–93	34 1961–93	33 1962–93
Aver. km^3/yr.	–143	–105	–428	181	–61	–297	–178	–149
Aver. km^3/yr.	–11	–26	–16	3	–4	–35	–90	–101

Data from Dyurgerov and Meier, 1997.
The glacier area and total mass balance include Svalbard glaciers.

the rate of carbon loss in Arctic ecosystems, providing a positive feedback for carbon dioxide and methane in the atmosphere (Oechel and Billings, 1992; Oechel et al., 1993; Reeburgh and Whalen, 1992; Zimov et al., 1997).

The United States Geological Survey has measured permafrost temperatures from deep drill holes in northern Alaska since the late 1940s. Based on data through the mid 1980s (Lachenbruch et al., 1982; Lachenbruch and Marshall, 1986), permafrost in this region generally warmed. Typical changes are 2 to 4 °C although some holes show little or no change or a cooling. The recent part of the records points to cooling in the early 1980s. Modeling studies of the penetration of the warming signal suggest that the warming began about 40 to 80 years ago. It has also been suggested that higher air temperatures in the late 1800s and early 1900s preceded the permafrost warming. However, Zhang and Osterkamp (1993) argue that at Barrow, air temperature variations alone (since 1923) cannot account for the observed warming, implicating changes in the snow or vegetation cover or perhaps an earlier warming.

Near-surface permafrost temperatures in northern Russia have also increased by 0.6–0.7 °C during the period 1970–1990 (Pavlov, 1994). Pavlov argues that this warming is more likely related to deeper snow cover rather than higher air temperatures although Figure 1 shows warming over northern Eurasia since 1966. By comparison, permafrost surface temperatures for northern Quebec appear to have decreased since the mid 1980s (Wang and Allard, 1995), tentatively attributed to lower air temperatures in this region (Figure 1).

Northern Alaskan data (1983 to 1993) reveal a cyclic variation in permafrost temperatures superimposed on the century-long warming and with similar amplitude (Osterkamp et al., 1994; Osterkamp and Romanovsky, 1996). Permafrost cooled initially until the mid 1980s, warmed until the early 1990s and then cooled until 1993, followed again by warming. This pattern is supported by other investigations for Alaska (Nelson et al., 1993; Lachenbruch, 1994). The two sites nearest the Arctic coast suggest a period of 10+ years, with an amplitude at the permafrost surface of about 2 °C. Two sites farther from the coast have similar periods but reduced amplitudes (1 °C). Kazantsev (1994) reports a similar behavior in Eastern Siberia and the Russian Far East and suggests a linkage with the solar cycle.

Active layer thickness variations have not generally followed those in permafrost temperature. Thicknesses on the Arctic Coastal Plain were largest in 1989 but in 1993, the warmest year of the decade, thicknesses ranged from average to near minimum (Romanovsky and Osterkamp, 1997). Thicknesses at West Dock, Prudhoe Bay, Alaska, have remained near minimum for the period from 1992 through 1996 while permafrost temperatures have risen. This is not entirely surprising as active layer thicknesses are determined primarily by summer conditions while permafrost temperatures reflect changes in mean annual conditions.

Alaskan data show that temperatures in discontinuous permafrost along a transect from Old Man near the Arctic Circle to Glenallen and at Healy generally increased in the late 1980s or early 1990s. This trend was not followed at Eagle

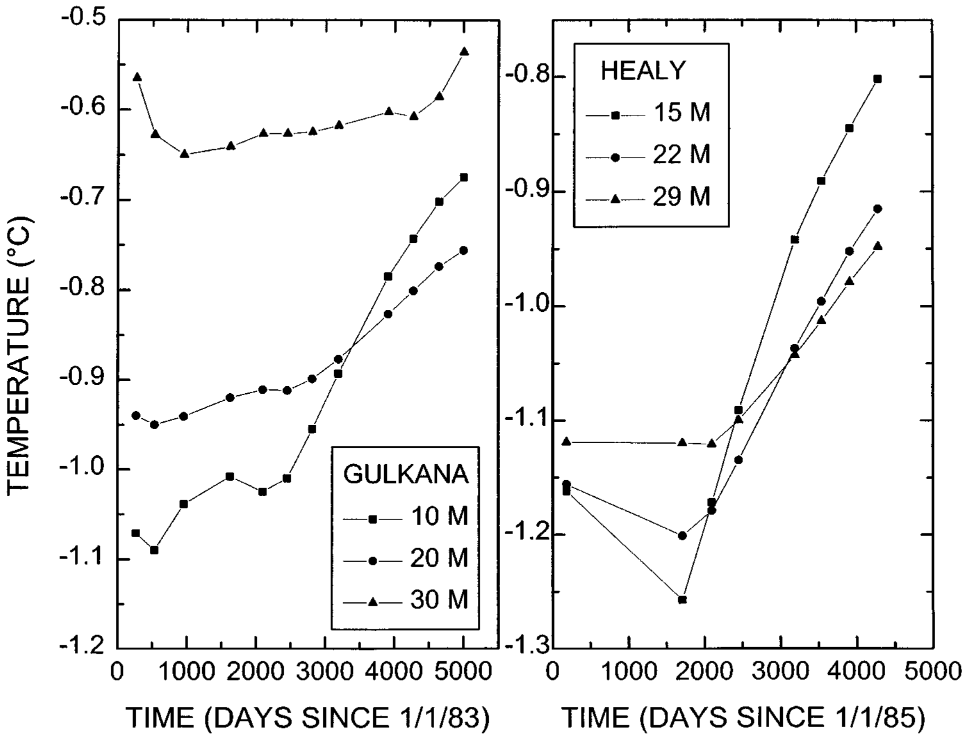

Figure 16. Temperature records in discontinuous permafrost near Gulkana and Healy, Alaska (from Osterkamp and Romanovsky, 1999).

which is about 330 km east of the transect. Examples of the warming at Healy and Gulkana are shown in Figure 16. Estimates for the magnitude of the warming at the permafrost table are typically 1.0–1.5 °C. Mean discontinuous permafrost temperatures in marginal areas were generally above –0.5 °C, indicating that the warming has probably caused permafrost in these areas to begin thawing. Thawing permafrost and thermokarst have been observed at several sites (Osterkamp, 1995; Osterkamp and Romanovsky, 1996). Warming of the permafrost in the last decade may result from fortuitous combinations of changes in snow cover thicknesses and air temperatures and thicker snow covers in the early 1990s according to Osterkamp and Romanovsky (1999).

5. Terrestrial Ecosystems

5.1. CARBON DIOXIDE AND METHANE FLUXES

The Arctic has been an overall significant sink for carbon over historic and recent geologic time scales, resulting in large stores of soil carbon of perhaps 300 gigatons (Miller et al., 1983). Carbon dating (^{14}C) of peat accumulation indicates carbon

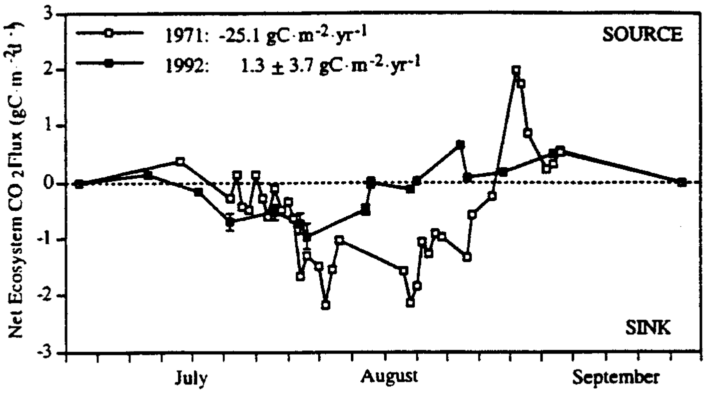

Figure 17. Seasonal net CO_2 flux measured during the 1971 and 1992 growing seasons at site IBP-II. Data are daily flux totals calculated over a 24-hour period. Positive values indicate a net atmospheric CO_2 source while negative values indicate a net sink (from Oechel et al., 1995).

uptake by Arctic terrestrial ecosystems on the north slope of Alaska through the Holocene (Marion and Oechel, 1993). Studies conducted under the International Biological Program (IBP) in the 1970s showed uptake rates of 30–100 g m^{-2} per year (Chapin et al., 1980; Miller et al., 1983). However, recent data suggest that past carbon accumulation has changed to a pattern of net loss, with growing season releases of up to 150 g m^{-2} y^{-1} (Marion and Oechel, 1993; Oechel et al., 1993; Zimov et al., 1993, 1996). These changes appear to represent significant deviations from historic and Holocene carbon fluxes, and the potential for a positive feedback on global change through losses of CO_2 to the atmosphere of up to 0.7 Gt C y^{-1} (about 12% of the total emission from fossil fuel use) (Oechel and Vourlitis, 1994).

To investigate this apparent change, net CO_2 fluxes measured at IGB site 2 in Barrow Alaska in 1971 (Coyne and Kelley, 1975) were reassessed in 1992. Both data sets comprise chamber and aerodynamic measurements. Data were also collected at surrounding sites at Barrow in the 1960s and first half of the 1990s (Oechel et al., 1995). The new measurements show that by 1992, the net CO_2 sinks of –25 g m^{-2} y^{-1} had become small sources of about 1 g m^{-2} y^{-1} (Figure 17). Even in years when the tundra at this site appears to be a sink for CO_2 during the growing season, it is a source when the full year is analyzed (Oechel et al., 1997). The results in Figure 17 are representative of wet sedge tundra of north Alaska, a common coastal vegetation type in Arctic regions. Wet sedge tundra accounts for about 18% of the circumpolar tundra (Oechel and Billings, 1992).

On a larger scale, the Kuparuk Basin (approximately 2.0×10^5 km^2) now appears as a net source of CO_2. This region is comprised mainly of acidic and non-acidic tussock tundra and wet sedge tundra (Oechel and Billings, 1992). Approximately 20% of the growing season loss is from carbon transported to lakes and streams in groundwater and then released from water sources to the atmosphere (Kling et al., 1991). Studies from Europe, Russia and Canada also show a preponderance of Arctic sites now losing carbon dioxide to the atmosphere (Zimov et al. 1993, 1996; Zamolodchikov and Karelin, unpublished). However, there are Arctic sites which are neutral or a sink for CO_2 (Sogaard et al., 1999).

Existing evidence suggests that the change in carbon flux to a small atmospheric source observed in Alaska is due to the effects of recent warming and resultant change in P-E on soil moisture content and soil water table and not to the direct effects of increasing temperature on ecosystem respiration. Drying has been shown to cause increased carbon loss in the Arctic under experimental conditions (Oechel et al., 1998) and drying has been observed in Barrow and the surrounding area (Oechel et al., 1995). Warming, where soil moisture is unchanged, would not be expected to cause a decrease in net ecosystem carbon sequestration (Shaver et al., 1992; Oechel and Vourlitis, 1994; Oechel et al., 1998). From Figures 2 and 3, it is apparent that since the early 1970s, there has been warming over Arctic land areas, largest in winter and spring. However, this has been attended by increases in annual precipitation, largely driven by winter. By comparison, summer and autumn precipitation since the early 1970s has remained fairly even, counter to the pattern of increase for the twentieth century as a whole. Annual P-E averaged for the polar cap shows no apparent trends since 1974 (Section 2.4), but we are not aware of any systematic studies focusing on land areas. Furthermore, it is apparent from Figure 1 that for Alaska, where most studies of the carbon flux have focused, warming has been particularly pronounced as compared to other regions of the Arctic. Clearly, there is a need for additional data on tundra carbon fluxes, coupled with more focused analyses of changes in the terrestrial hydrologic budget, and the seasonality and spatial extent of these changes.

Thermokarst, which is expected to increase in response to observed warming of permafrost, could increase methane fluxes by increasing the area of wetlands and ponds. High-latitude wetlands currently account for 5–10% of global fluxes of methane (Reeburgh and Whalen, 1992). In addition, Siberian thermokarst lakes, which emit most of their methane in winter, could contribute to the recent increase in seasonal amplitude and winter concentration of atmospheric methane observed at high latitudes (Zimov et al., 1997). Methane release from thermokarst lakes is fueled primarily by Pleistocene carbon of terrestrial origin. However, the time series of methane release are too short to detect trends (Whalen and Reeburgh, 1992).

5.2. VEGETATION CHANGES

Mynemi et al. (1997) present evidence that photosynthetic activity of terrestrial vegetation in northern high latitudes increased from 1981 through 1991 suggestive of an increase in plant growth and a lengthening of the active growing season. The largest increases in photosynthetic activity (10–12%) are found between 45–70° N, which they argue is consistent with marked springtime warming. Results are based on two independent records of the normalized difference vegetation index (NDVI) derived from NOAA Advanced Very High Resolution Radiometer (AVHRR) satellite records. Further analyses show continuation of the increasing NDVI on the north slope of Alaska into 1997 (Hope et al., unpublished).

Results appear consistent with an increased amplitude in the seasonal cycle of atmospheric carbon dioxide of over 20% since the 1970s at Point Barrow, Alaska, and an advance of up to seven days in the timing of CO_2 drawdown in spring and early summer (Keeling et al., 1996). Fung (1997), in arguing in general for the veracity of these results, points out that over the same period, CO_2 has increased by only 4% (from 340–355 ppmv) and could not have enhanced photosynthesis at the NVDI rate. In addition to the possibility that temperature increases may have stimulated photosynthesis directly or indirectly by accelerating snowmelt and increasing the length of the growing season, Fung (1997) also argues that higher temperatures may have mobilized nutrients previously frozen in the soil. The NDVI record is obviously too short to make firm conclusions. In this regard, Jones and Briffa (1995) find that over the FSU, there have been no coherent changes in the duration, start or end of the growing season for the period 1950–1989 (and since the 1880s for selected stations with long temperature records). This is consistent with observations that temperature increases have been strongest for the winter season.

While interpretation of the NDVI time series is open to debate, observations do point to a northward movement of the Arctic tree line in recent decades (D’Arrigo et al., 1987; Nichols, 1998). The tree line is closely associated with the July position of the Arctic front. While the front represents a climatic forcing on the position of the tree line, with colder conditions to the north providing conditions unsuitable for trees to establish, discontinuities in energy exchange between forest and tundra may also help to stabilize the front at this location (Bryson, 1966; Krebs and Barry, 1970; Pielke and Vidale, 1996). The 1980s and 1990s have also seen an increased abundance of shrubs in northern Alaska (Chapin et al., 1995). The paleoclimate record (Brubaker et al., 1995) indicates similar changes during previous Holocene warming events.

There have also been increases in fire frequency in Alaska between 1955 and 1992 (Oechel and Vourlitis, 1996) and in other circumpolar zones that have experienced warming (Stocks, 1991), but whether these changes are climate induced or a result of reckless human behavior has not been resolved. Fire can accelerate the loss of carbon as CO_2 to the atmosphere during and after the fire. In addition, fire can facilitate the conversion of previously unforested subarctic tundra systems to forested boreal ecosystems by removing the insulating surface organic layer and exposing mineral soil, promoting permafrost thawing and providing conditions that favor the germination and establishment of some boreal tree species (Landhausser and Wein, 1993; Oechel and Vourlitis, 1996; Mackay, 1995).

6. Conclusions

Appendix 1 summarizes observations for the primary variables discussed in this paper. On the balance, the records are compatible with each other, painting a reasonably coherent picture of recent environmental change in northern high latitudes. Air temperature increases over northwest North America and Eurasia, largest for winter and spring, have been accompanied by spring and summer warming over the Arctic Ocean. Reconstructions from proxy sources imply that Arctic air temperatures in the 20th century are the highest in the past 400 years. Satellite records point to a slight downward trend in sea ice extent and a longer melt season while other data indicate warming and increased areal extent of the Arctic Ocean's Atlantic layer. Negative snow cover anomalies have characterized both North America and Eurasia since the late 1980s, primarily during spring and summer, and precipitation has increased over northern high latitudes since 1900. The mass balance of small Arctic glaciers has been generally negative, paralleling a global tendency, and there is evidence of increasing permafrost temperatures. Changes in vegetation and trace gases fluxes from the tundra are also consistent with warming.

These conclusions, however, are subject to a number of caveats. The time periods examined for the different variables vary widely and some (e.g., the sea ice and most ecological records) are short, presenting problems in assessing consistency between trends. The data sources also differ markedly in their ability to assess spatial characteristics of change. For example, glacier mass balance, permafrost temperature and tundra carbon flux records are available only from a limited number of sites. The quality of some time series (e.g., precipitation) is also suspect. Assessing the significance of trends is also complicated by knowledge of and assumptions regarding the temporal autocorrelation structure of the residuals about a trend.

Do the results constitute evidence of an anthropogenic influence on climate? On the one hand, the pattern of temperature change observed over the past few decades (hence change in many other variables) is consistent with the persistence of modes of atmospheric circulation, including the AO, NAO, and extratropical responses to tropical SST forcing. In turn, the recent warming, while pronounced, is no larger than the observed interdecadal range in high-latitude temperatures during this century. On the other hand, an accounting of the effects of atmospheric circulation as well as forcing by insolation and volcanic aerosols (Hurrell, 1996; Wallace et al., 1996; Overpeck et al., 1997) leaves a residual warming and the general spatial patterns of recent temperature change agree with model results. Furthermore, it is reasonable to expect that a radiatively-induced background rise in global temperature could shift the configuration of the planetary longwaves to favor persistent circulation modes that enhance high-latitude warming. Recent modeling experiments (e.g., Broccoli et al., 1998; Osborn et al., 1999, Fyfe et al., 1999) indicate that anthropogenic forcing may indeed modulate the intensity and frequency of modes of variability such as the NAO, AO and ENSO and contribute to the COWL pattern of temperature change.

It is clear that effective monitoring of the Arctic climate requires improvements in observational data bases. Unfortunately, this requirement is contrary to the closures of various hydrometeorological stations and reductions of other networks in Canada and in Russia where the North Pole drifting program was terminated in 1991 (Barry, 1995). A new reanalysis by ECMWF is expected to provide improved long-term records for the Arctic, but will still contain temporal inhomogeneities related to changes in the amount and quality of assimilation data. Reliable long-term time series of other important variables, such as Arctic cloud cover, do not exist. We advocate continued efforts into the development of gridded data bases of Arctic cloud properties and surface radiation fluxes from satellite data using improved retrieval algorithms, and efforts to update station data sets of surface radiation fluxes, precipitation and temperature. Continuance of existing programs providing valuable high-latitude data, such as the International Arctic Buoy Program must be assured. Network reductions should take careful account of their impacts on our ability to monitor for climate change.

Acknowledgements

This study was supported by NSF grants OPP-9321547, OPP-9614297, OPP-9634289, OPP-9530782, ATM-9315351, ATM-9314721, OPP-9504201, SBR-9320786, ATM-9319952 and OPP-9732461. The anonymous reviewers as well as M. Wallace and M. Steele are thanked for their constructive comments. We thank J. Hurrell for Figure 6, J. Overpeck and D. Thompson for the data used in Figures 5 and 8, respectively, D. Robinson for Figure 10 and J. Eischeid for Figures 3 and 11.

Appendix 1. Summary of Evidence for Change in Northern High Latitudes. Record Periods Examined in Major Studies Are Shown in Parentheses

Air Temperature (1966–1995; 1900–1995; 1979–1995; 1600–1990 from Proxy Sources)

Positive trends since 1966 over northern Eurasia and North America, largest during winter and spring with partly-compensating negative trends over eastern Canada, southern Greenland and the northern north Atlantic. Positive spring and summer trends over the Arctic Ocean since at least 1961. Reconstructions from proxy sources suggest that the mid-20th century Arctic warming is unprecedented over the past 400 years.

Atmospheric Circulation (1966–1993; 1979–1994; 1958–1997; 1947–1995; 1899–1997)
Generally positive phase of the NAO and of the AO since about 1970. Increased cyclone activity since at least 1958 north of 60° N for all seasons except autumn; increased cyclone intensity for all seasons. Pronounced increases at higher latitudes during summer. Reductions in central Arctic sea level pressure for both the cold and warm seasons. Circulation changes are consistent with observed temperature trends.

Precipitation (1900–1995)
General increases for the 55–85° N latitude band, largest during autumn and winter. Pronounced recent increases in the past 40 years over northern Canada.

Precipitation Minus Evaporation North of 70° N (1974–1991; 1973–1996)
No apparent trends.

Snow Covered Area and Snow Depth (1972–1997; 1946–1995; 1891–1992; 1915–1992 from Reconstructions)
Satellite records indicate a decrease in Northern Hemisphere annual snow covered area (SCA) of about 10% since 1972, largely reflecting spring and summer deficits since the mid 1980s. Station records indicate a general decrease in snow depth since 1946 over Canada, especially during spring, and depths have declined over European Russia since the turn of the century. Reconstructions for Canada suggest a general decrease in spring SCA since 1915 but increased winter SCA. Winter snow depths over parts of Russia also appear to have increased in recent decades. A common thread between studies examining seasonality is a reduction in spring snow cover.

Sea Ice Extent (1961–1990; 1979–1996; 1900–Present; 1500s–Present)
Passive microwave record indicates a small negative trend since 1979 with more pronounced reductions since the late 1980s, dominated by negative anomalies in late summer along the Siberian coast. Record low ice cover in the western Arctic in 1998. Increase in the length of the melt season. Increased sea ice extent in Antarctica. Regional *in situ* data sources suggest reductions during the twentieth century in the eastern North Atlantic.

Ocean Structure (Differences between 1970s to 1980s and Early to Mid 1990s)
Increased warming of the Atlantic layer and 20% greater coverage of Atlantic water types. The change may have started in late 1980s.

Permafrost (Variable, Longest Records Starting in 1950s)
Permafrost temperature increases for Alaska and Siberia but not consistent. Decreased temperatures for eastern Canada. Poor spatial coverage of observations.

Glacier Mass Balance (Variable, Earliest Records Starting in 1940s)
Unknown for Greenland. Generally negative cumulative balances for small glaciers over the Arctic as a whole, Canada, Svalbard and Alaska. The Arctic appears to account for about 20% of the estimated 7.4 mm global sea level rise since 1961 due to melt of small glaciers.

Plant Growth: (1981–1991)
Increased NDVI in 45–70° N band (based on satellite data) suggestive of increased plant growth, but the record length is short. Northward migration of tree line over the past several decades and expansion of shrubs in Alaska.

Carbon Flux (Differences between Measurements in Early 1970s and Early 1990s)
Change from tundra from a net sink of atmospheric CO_2 to a small net source. Poor spatial coverage of observations.

References

Aagaard, K. and Carmack, E. C.: 1989, 'The Role of Sea Ice and Other Fresh Waters in the Arctic Circulation', *J. Geophys. Res.* **94** (C10), 14,845–14,498.

Abdalati, W. and Steffen, K.: 1997, 'Snowmelt on the Greenland Ice Sheet as Derived from Passive Microwave Data', *J. Climate* **10**, 165–175.

Anderson, L. G., Bjork, B., Holby, O., Jones, E. P., Kattner, G., Koltermann, K. P., Liljeblad, B., Lindegren, R., Rudels, B., and Swift, J. H.: 1994, 'Water Masses and Circulation in the Eurasian Basins: Results from the Oden 91 Expedition', *J. Geophys. Res.* **99** (C2), 3273–3283.

Angell, K. and Korshover, J.: 1974, 'Quasi-Biennial and Long-Term Fluctuations in Centers of Action', *Mon. Wea. Rev.* **102**, 669–678.

Barry, R. G: 1995, 'Observing Systems and Data Sets Related to the Cryosphere in Canada: A Contribution to Planning for the Global Climate Observing System, *Atmos.-Ocean* **33**, 771–807.

Barry, R. G. and Perry, A. H.: 1973, *Synoptic Climatology. Methods and Applications*, Methuen, London, pp. 161–165.

Basist, A. N. and Chelliah, M.: 1997, 'Comparison of Tropospheric Temperatures Derived from the NCEP/NCAR Reanalysis, NCEP Operational Analysis and the Microwave Sounding Unit', *Bull. Amer. Meteor. Soc.* **7**, 1431–1447.

Billings, W. D., Luken, J. O., Martensen, D. A., and Peterson, K. M.: 1983, 'Increasing Atmospheric Carbon Dioxide: Possible Effects on Arctic Tundra', *Oecologia* **58**, 286–289.

Bjorgo, E., Johannessen, O. M., and Miles, M. W.: 1997, 'Analysis of Merged SSMR-SSMI Time Series of Arctic and Antarctic Sea Ice Parameters 1978–1995', *Geophys. Res. Lett.* **24**, 413–416.

Broccoli, A. J., Lau, N.-C., and Nath, M. J.: 1998, 'The Cold Ocean–Warm Land Pattern: Model Simulation and Relevance to Climate Change Detection', *J. Climate* **11**, 2743–2763.

Brown, R. D. and Braaten, R. O.: 1998, 'Spatial and Temporal Variability of Canadian Monthly Snow Depths, 1946–1995', *Atmos.–Ocean* **36**, 37–54.

Brown, R. D. and Goodison, B. E.: 1996, 'Interannual Variability in Reconstructed Canadian Snow Cover, 1915–1992', *J. Climate* **9**, 1299–1318.

Brubaker, L. B., Anderson, P. M., and Hu, F. S.: 1995, 'Arctic Tundra Biodiversity: A Temporal Perspective from Late Quaternary Pollen Records', in Chapin, F. S. III and Kšrner, C. (eds.), *Arctic and Alpine Biodiversity: Patterns, Causes and Ecosystem Consequences*, Springer-Verlag, Berlin, pp. 111–125.

Bryson, R. A.: 1966, 'Air Masses, Stream Lines, and the Boreal Forest', *Geogr. Bull.* **8**, 228–269.

Carmack, E. C., Aagaard, K, Swift, J. H., Macdonald, R. W., McLaughlin, F. A., Jones, E. P., Perkin, R. G., Smith, J. N., Ellis, K., and Kilius, L.: 1998, 'Changes in Temperature and Contaminant Distributions within the Arctic Ocean', *Deep Sea Res. Par II* **44**, 1487–1502.

Cavalieri, D. J, Gloersen, P, Parkinson, C. L., Comiso, J. C., and Zwally, H. J.: 1997, 'Observed Hemispheric Asymmetry in Global Sea Ice Changes', *Science* **278**, 1104–1106.

Chapin, F. S. III, Miller, P. C., Billings, W. D., and Coyne, P. I.: 1980, 'Carbon and Nutrient Budgets and their Control in Coastal Tundra', in Brown, J., Miller, P. C., Tieszen, L. L., and Bunnell, F. L. (eds.), *An Arctic Ecosystem, the Coastal Tundra at Barrow, Alaska*, Dowden, Hutchinson and Ross, Stroudsberg, PA, pp. 458–482.

Chapin, F. S. III, Shaver, G. R., Giblin, A. E., Nadelhoffer, K. G., and Laundre, J. A.: 1995, 'Response of Arctic Tundra to Experimental and Observed Changes in Climate', *Ecology* **76**, 694–711.

Chapman, W. L. and Walsh, J. E.: 1993, 'Recent Variations of Sea Ice and Air Temperature in High Latitudes', *Bull. Amer. Meteorol. Soc.* **74**, 33–47.

Colony, R. L. and Rigor, I. G.: 1993, 'International Arctic Ocean Buoy Program Data Report for 1 January 1992–31 December 1992', *Tech. Mem. APL-UWTM 29-93, APPL*, Univ. Washington, Seattle WA, p. 215.

Colony, R. L., Radionov, V., and Tanis, F. L.: 1997, 'Measurements of Precipitation and Snow at the Russian North Pole Drifting Stations', *Polar Geogr.* **34**, 3–14.

Coyne, P. I. and Kelley, J. J.: 1975, 'CO_2 Exchange over the Alaskan Tundra: Meteorological Assessment by an Aerodynamic Method', *J. Appl. Ecol.* **12**, 587–610.

Cullather, R. I., Bromwich, D. H., and Serreze, M. C.: 2000, 'The Atmospheric Hydrologic Cycle over the Arctic Basin from Reanalyses Part I. Comparison with Observations and Previous Studies', *J. Climate*, in press.

Dai, A., Fung, I. Y., and Del Genio, A. D.: 1997, 'Surface Observed Global Land Precipitation Variations during 1900–1988', *J. Climate* **10**, 2943–2962.

D'Arrigo, R., Jacoby, G. C., and Fung, I. Y.: 1987, 'Boreal Forests and Atmosphere–Biosphere Exchange of Carbon Dioxide', *Nature* **329**, 321–323.

Dickson, R. R., Osborn, T. J., Hurrell, J. W., Meincke, J., Blindheim, J., Adlandsvik, B., Vinje, T., Alekseev, G., Maslowski, W., and Cattle, H.: 2000, 'The Arctic Ocean Response to the North Atlantic Oscillation', *J. Climate*, in press.

Dmitriev, A. A.: 1994, *Iznerchivost Atmosfernykh Protsessy Arktiki i ee Ychet v Dolgosrochnykh Prognozakh (Variability of Atmospheric Processes in the Arctic and their Calculation for Long-Range Prediction)*, Gidrometeoizdat, St. Petersburg, p. 207.

Dowdeswell, J. A., Hagen, J. O., Bjornsson, H., Glazovsky, A. F., Harrison, W. D., Holmlund, P., Jania, J., Koerner, R. M., Lefauconnier, B., Ommanney, C. S. L., and Thomas, R. H.: 1997, 'The Mass Balance of Circum-Arctic Glaciers and Recent Climate Change', *Quatern. Res.* **48**, 1–14.

Dyurgerov, M. B. and Meier, M. F.: 1997, 'Year-to-Year Fluctuation of Global Mass Balance of Small Glaciers and their Contribution to Sea Level Changes', *Arc. Alp. Res.* **29**, 392–402.

Dzerdzeevskii, B. L.: 1963, 'Fluctuations of General Circulation of the Atmosphere and Climate in the Twentieth Century', in *Changes of Climate*, Arid Zone Res. No. 20, UNESCO, Paris, pp. 285–295.

Eischeid, J. K., Baker, C. B., Karl, T. R., and Diaz, H. F.: 1995, 'The Quality Control of Long-Term Climatological Data Using Objective Data Analysis', *J. Appl. Meteorol.* **34**, 2787–2795.

Environmental Working Group (EWG): 1998, *Joint U.S. – Russian Atlas of the Arctic Ocean, Oceanography Atlas for the Winter Period*, National Ocean Data Center, CD-ROM available from the National Snow and Ice Data Center (NSIDC), Boulder, CO.

Fallot, J.-M., Barry, R. G., and Hoogstrate, D.: 1997, 'Variations of Mean Cold Season Temperature, Precipitation and Snow Depths during the Last 100 Years in the Former Soviet Union (FSU)', *Hydrol. Sci. J.* **42**, 301–327.

Fung, I.: 1997, 'A Greener North', *Nature* **386**, 659–660.

Fyfe, J. C., Boer, G. J., and Flato, G. M.: 1999, 'The Arctic and Antarctic Oscillations under Global Warming', *Geophys. Res. Lett.* **26**, 1601–1604.

Girs, A. A.: 1981, 'K Voprosy o Formakh Atmosfernoi Tsirkulyatsii i ikh Pragnosticheskom Izpolzobanii (On the Forms of Atmospheric Circulation and their Use in Forecasting)', *Trudy Arkt. Antarkt. Inst.* **373**, 4–13.

Gloersen, P. and Campbell, W. J.: 1991, 'Recent Variations in Arctic and Antarctic Sea-Ice Covers', *Nature* **352**, 33–36.

Groisman, P. Y. and Easterling, D. R.: 1994, 'Variability and Trends of Precipitation and Snowfall over the Eastern United States and Canada', *J. Climate* **7**, 184–205.

Groisman, P. Y., Karl, T. R., Knight, R. W., and Stenchinkov, G. L.: 1994a, 'Changes of Snow Cover, Temperature and Radiative Heat Balance over the Northern Hemisphere', *J. Climate* **7**, 1633–1656.

Groisman, P. Y., Karl, T. R., and Knight, T. W.: 1994b, 'Observed Impact of Snow Cover on the Heat Balance and the Rise of Continental Spring Temperatures', *Science* **263**, 198–200.

Groisman, P. Y., Koknaeva, V. V., Belokrylova, T. A., and Karl, T. R.: 1991, 'Overcoming Biases of Precipitation: A History of the U.S.S.R. Experience', *Bull. Amer. Meteorol. Soc.* **72**, 1725–1733.

Grotefendt, K., Logemann, K., Quadfasel, D., and Ronski, S.: 1998, 'Is the Arctic Ocean Warming?', *J. Geophys. Res.* **103** (C12), 27,679–27,687.

Hurrell, J. W.: 1995, 'Decadal Trends in the North Atlantic Oscillation: Regional Temperatures and Precipitation', *Science* **269**, 676–679.

Hurrell, J. W.: 1996, 'Influence of Variations in Extratropical Wintertime Teleconnections on Northern Hemisphere Temperature', *Geophys. Res. Lett.* **23**, 665–668.

Intergovernmental Panel of Climate Change (IPCC): 1992, in Houghton, J. T., Callander, B. A., and Varney, S. K. (eds.), *Climate Change 1992: The Supplementary Report to the IPCC Scientific Assessment*, Cambridge University Press, Cambridge, U.K., p. 200.

Johannessen, O. M., Miles, M., and Bjorgo, E.: 1995, 'The Arctics Shrinking Sea Ice', *Nature* **376**, 126–127.

Jones, P. D.: 1994, 'Hemispheric Surface Air Temperature Variations: A Reanalysis and an Update to 1993', *J. Climate* **7**, 1794–1802.

Jones, P. D. and Briffa, K. R.: 1995, 'Growing Season Temperatures over the Former Soviet Union', *Int. J. Climatol.* **15**, 943–959.

Kahl, J. D., Charlevoix, D. J., Zaitseva, N. A., Schnell, R. C., and Serreze, M. C.: 1993, 'Absence of Evidence for Greenhouse Warming over the Arctic Ocean in the Past 40 Years', *Nature* **361**, 335–337.

Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woolen, J., Zhu, Y., Chelliah, M., Ebisuzaki, W., Higgens, W., Janowiak, J., Mo, K. C., Ropelewski, C., Wang, J., Leetma, A., Reynolds, R., Jenne, R., and Joseph, D.: 1996, 'The NCEP/NCAR 40-Year Reanalysis Project', *Bull. Amer. Meteorol. Soc.* **77**, 437–471.

Kattenberg, A., Giorgi, F., Grassl, H., Meehl, G. A., Mitchell, J. F. B., Stouffer, R. J., Tokioka, T., Weaver, A. J., and Wigley, T. M. L.: 1996, 'Climate Models–Projections of Future Climate', in Houghton, J. T., Meira Filho, L. G., Callander, B. A., Harris, N., Kattenberg, A., and Maskell, K. (eds.), *Climate Change 1995, The Science of Climate Change, Contribution of Working Group I to the Second Assessment of the Intergovernmental Panel on Climate Change* (Ch. 6), Cambridge University Press, Cambridge, U.K., pp. 289–357.

Kazantsev, V. V.: 1994, 'Is the Thermal Regime of Permafrost Determined by Solar Rhythms?', *Cold Regions Sci. Technol.* **23**, 93–98.

Keeling, C. D., Chin, J. F. S., and Whorf, T. P.: 1996, 'Increased Activity on Northern Vegetation Inferred from Atmospheric CO_2 Measurements', *Nature* **382**, 146–149.

Kling, G. W., Khiyshut, G. W., and Miller, M. C.: 1991, 'Arctic Lakes and Streams as Gas Conduits to the Atmosphere: Implications for Tundra Carbon Budgets', *Science* **251**, 298–301.

Koch, L.: 1945, 'The East Greenland Ice', *Medd. Groenl.* **130**, 1–374.

Krebs, S. J. and Barry, R. G.: 1970, 'The Arctic Front and the Tundra–Taiga Boundary in Eurasia', *Geog. Rev.* **60**, 548–554.

Lachenbruch, A. H.: 1994, 'Permafrost, the Active Layer, and Changing Climate', United States Geological Survey, Open-File Report 94-694, p. 43.

Lachenbruch, A. H. and Marshall, B. V.: 1986, 'Changing Climate: Geothermal Evidence from Permafrost in the Alaskan Arctic', *Science* **234**, 689–696.

Lachenbruch, A. H., Sass, J. H., Marshall, B. V., and Moses, T. H.: 1982, 'Permafrost, Heat Flow, and the Geothermal Regime at Prudhoe Bay, Alaska', *J. Geophys. Res.* **87** (B11), 9301–9316.

Landhausser, S. M. and Wein, R. W.: 1993, 'Postfire Vegetation Recovery and Tree Establishment at the Arctic Treeline: Climate-Change Vegetation-Regeneration Hypotheses', *J. Ecol.* **81**, 665–672.

Lean, J., Beer, J., and Bradley, R.: 1995, 'Reconstruction of Solar Irradiance Since 1610: Implications for Climate Change', *Geophys. Res. Lett.* **22**, 3195–3198.

Mächel, H., Kapala, A., and Flohn, H.: 1998, 'Behaviour of the Centres of Action Above the Atlantic Since 1881. Part I: Characteristics of Seasonal and Interannual Variability', *Int. J. Climatol.* **18**, 1–22.

Mackay, J. R.: 1995, 'Active Layer Changes (1968–1993) Following the Forest-Tundra Fire near Inuvik, N.W.T., Canada', *Arc. Alp. Res.* **27**, 323–336.

Manabe, S., Spelman, M. J., and Stouffer, R. J.: 1992, 'Transient Response of a Coupled Ocean-Atmosphere Model to Gradual Changes of Atmospheric CO_2. Part II: Seasonal Response', *J. Climate* **5**, 105–126.

Marion, G. M. and Oechel, W. C.: 1993, 'Mid-to-Late-Holocene Carbon Balance in Arctic Alaska and its Implications for Future Global Warming', *Holocene* **3**, 193–200.

Martin, S., Munoz, E., and Dreucker, R.: 1997, 'Recent Observations of a Spring–Summer Warming over the Arctic Ocean', *Geophys. Res. Lett.* **24**, 1259–1262.

Martin, S. E. and Munoz, E.: 1997, 'Properties of the Arctic 2-m Air Temperature for 1979–Present Derived from a New Gridded Data Set', *J. Climate* **10**, 1428–1440.

Maslanik, J. A., Serreze, M. C., and Agnew, T.: 1999, 'On the Record Reduction in Western Arctic Sea Ice Cover in 1998', *Geophys. Res. Lett.* **26**, 1905–1908.

Maslanik, J. A., Serreze, M. C., and Barry, R. G.: 1996, 'Recent Decreases in Arctic Summer Ice Cover and Linkages to Atmospheric Circulation Anomalies', *Geophys. Res. Lett.* **23**, 1677–1680.

McLaughlin, F. A., Carmack, E. C., McDonald, R. W., and Bishop, J. K. B.: 1996, 'Physical and Geochemical Properties Across the Atlantic/Pacific Water Mass Front in the Southern Canada Basin', *J. Geophys. Res.* **101** (C1), 1183–1195.

Meehl, G. A. and Washington, W. M.: 1990, 'CO_2 Climate Sensitivity and Snow-Sea-Ice Parameterization in an Atmospheric GCM Coupled to a Mixed-Layer Ocean Model', *Clim. Change* **16**, 283–306.

Meehl, G. A., Boer, G. J., Covey, C., Latif, M., and Stouffer, R. J.: 1997, 'Intercomparison Makes for a Better Climate Model, *EOS* **78**, 445–451.

Meier, M. F.: 1984, 'Contribution of Small Glaciers to Global Sea Level', *Science* **226**, 1418–1421.

Meshcherskaya, A. V., Belyankina, I. G., and Golod, M. P.: 1995, 'Monitoring Tolshching Cnozhnogo Pokprova v Osnovioi Zerno Proizvodyashchei Zone Byvshego SSSR za Period Instrumentalnykh Nablyugenii', *Izvestiya Akad. Nauk SSR., Sser. Geograf.*, 101–110.

Mikhalevesky, P. N., Gavrilov, A., Baggoroer, A. P., and Slavinsky, M.: 1995, 'Experiment Tests Use of Acoustics to Monitor Temperature and Ice in the Arctic Ocean', *Eos, Trans. AGU* **76**, 27.

Miller, P. C., Kendall, R., and Oechel, W. C.: 1983, 'Simulating Carbon Accumulation in Northern Ecosystems', *Simulation* **40**, 119–131.

Morison, J. H., Steele, M., and Andersen, R.: 1998, 'Hydrography in Upper Arctic Ocean Measured from the Nuclear Submarine U.S.S. Pargo', *Deep Sea Res. Part 1* **45**, 15–38.

Myneni, R. B., Keeling, C. D., Tucker, C. J., Asrar, G., and Nemani, R. R.: 1997, 'Increased Plant Growth in the Northern High Latitudes from 1981 to 1991', *Nature* **386**, 698–702.

Nakamura, N. and Oort, A. H.: 1988, 'Atmospheric Heat Budgets of the Polar Regions', *J. Geophys. Res.* **93** (D8), 9510–9524.

Nelson, F. E., Lachenbruch, A. H., Woo, M. K., Koster, E. A., Osterkamp, T. E., Gavrilova, M. K., and Cheng, G. D.: 1993, 'Permafrost and Climate Change', in *Proceedings of the Sixth International Conference on Permafrost,* Vol. 2, South China University of Technology Press, pp. 987–1005.

Nicholls, N., Gruza, G. V., Jouzel, J., Karl, T. R., Ogallo, L. A., and Parker, D. E.: 1996, 'Observed Climate Variability and Change', in Houghton, J. T., Meira Filho, L. G., Callander, B. A., Harris, N., Kattenberg, A., and Maskell, K. (eds.), *Climate Change 1995, The Science of Climate Change, Contribution of Working Group I to the Second Assessment of the Intergovernmental Panel on Climate Change* (Ch. 3), Cambridge University Press, Cambridge, U.K., pp. 137–192.

Oechel, W. C. and Billings, W. D.: 1992, 'Anticipated Effects of Global Change on Carbon Balance of Arctic Plants and Ecosystems', in Chapin, F. S. III, Jefferies, R., Shaver, G., Reynolds, J., and Svoboda, J. (eds.), *Physiological Ecology of Arctic Plants: Implications for Climate Change,* Academy Press, New York, NY, pp. 139–168.

Oechel, W. C. and Vourlitis, G. L.: 1994, 'The Effects of Climate Change on Arctic Tundra Ecosystems', *Trends Ecol. Evol.* **9**, 324–329.

Oechel, W. C. and Vourlitis, G. L.: 1996, 'Climate Change in Northern Latitudes: Alterations in Ecosystem Structure and Function and Effects on Carbon Sequestration', in Oechel, W. C., Callaghan, T., Gilmanov, T., Holten, J. I., Maxwell, B., Molau, U., and Sveinbjornsson, B. (eds.), *Global Change and Arctic Terrestrial Ecosystems* (Ch. 20), Springer, pp. 381–401.

Oechel, W. C., Hastings, S. J., Vourlitis, G., Jenkins, M., Richers, G., Gruike, N.: 1993, 'Recent Change of Arctic Tundra Ecosystems from a Net Carbon Dioxide Sink to a Source', *Nature* **361**, 520–523.

Oechel, W. C., Vourlitis, G. L., and Hastings, S. J.: 1997, 'Cold-Season CO_2 Emission from Arctic Soils', *Global Biogeochem. Cycles* **11**, 163–172.

Oechel, W. C., Vourlitis, G. L., Hastings, S. J., Ault, R. P. Jr., and Bryant, P.: 1998, 'The Effects of Water Table Manipulation and Elevated Temperature on the Net CO_2 Flux of Wet Sedge Tundra Ecosystems', *Global Change Biol.* **4**, 77–90.

Oechel, W. C., Vourlitis, G. L., Hastings, S. J., and Bochkarev, S. A.: 1995, 'Effects of Arctic CO_2 Flux over Two Decades: Effects of Climate Change at Barrow, Alaska', *Ecol. Appl.* **5**, 846–855.

Osborn, T. J., Briffa, K. R., Tett, S. F. B., Jones, P. D., and Trigo, R. M.: 1999, 'Evolution of the North Atlantic Oscillation as Simulated by a Climate Model', *Clim. Dyn.* **15**, 685–702.

Osterkamp, T. E.: 1995, 'Further Evidence for Warming and Thawing of Permafrost in Alaska', *EOS, Trans. AGU* **76**, F244.

Osterkamp, T. E. and Romanovsky, V. E.: 1996, 'Characteristics of Changing Permafrost Temperatures in the Alaskan Arctic', *Arc. Alp. Res.* **28**, 267–273.

Osterkamp, T. E. and Romanovsky, V. E.: 1999, 'Evidence for Warming and Thawing of Discontinuous Permafrost in Alaska', *Permafrost Periglacial Proc.* **10**, 17–37.

Osterkamp, T. E., Zhang, T., and Romanovsky, V. E.: 1994, 'Evidence for a Cyclic Variation of Permafrost Temperatures in Northern Alaska', *Permafrost Periglacial Proc.* **5**, 137–144.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A., and Zielinski, G.: 1997, 'Arctic Environmental Change of the Last Four Centuries', *Science* **278**, 1251–1256.

Parkinson, C. L. and Cavalieri, D. J.: 1989, 'Arctic Sea Ice 1973–1987: Seasonal, Regional and Interannual Variability', *J. Geophys. Res.* **94** (C10), 14,499–14,523.

Parkinson, C. L., Comiso, J., Zwally, H. J., Cavalieri, D. J., Gloersen, P., and Campbell, W. J.: 1987, 'Arctic Sea Ice, 1973–1976: Satellite Passive Microwave Observations', NASA Special Publication, SP-489, p. 295.

Pavlov, A. V.: 1994, 'Current Changes of Climate and Permafrost in the Arctic and Sub-Arctic of Russia', *Permafrost Periglacial Proc.* **5**, 101–110.

Pielke, R. A. and Vidale, P. L.: 1996, 'The Boreal Forest and the Polar Front', *J. Geophys. Res.* **100** (D), 25755–25758.

Quadfasel, D.: 1991, 'Warming in the Arctic', *Nature* **350**, 385.

Reeburgh, W. S. and Whalen, S. C.: 1992, 'High Latitude Ecosystems as CH_4 Sources, *Ecol. Bull.* **42**, 62–70.

Rind, D., Martinson, D., Parkinson, C., and Healy, R.: 1996, 'Sea Ice Forcing of Climate Change', in Workshop on Polar Processes in Global Change, Cancun, Mexico, American Meteorological Society, pp. 4–7.

Robinson, D. A., Dewey, K. F., and Heim, R. R.: 1993, 'Global Snow Cover Monitoring: An Update', *Bull. Amer. Meteorol. Soc.* **74**, 1689–1696.

Robinson, D. A., Frei, A., and Serreze, M. C.: 1995, 'Recent Variations and Regional Relationships in Northern Hemisphere Snow Cover', *Ann. Glaciol.* **21**, 71–76.

Robinson, D. A., Serreze, M. C., Barry, R. G., Scharfen, G., and Kukla, G.: 1992, 'Large-Scale Patterns and Variability of Snowmelt and Parameterized Surface Albedo in the Arctic Basin', *J. Climate* **5**, 1109–1119.

Rogers, J. C. and Mosley-Thompson, E.: 1995, 'Atlantic Arctic Cyclones and the Mild Siberian Winters of the 1980s', *Geophys. Res. Lett.* **22**, 799–802.

Romanovsky, V. E. and Osterkamp, T. E.: 1997, 'Thawing of the Active Layer on the Coastal Plain of the Alaskan Arctic', *Permafrost Periglacial Proc.* **8**, 1–22.

Rossow, W. B. and Schiffer, R. A.: 1991, 'ISCCP Cloud Data Products', *Bull. Amer. Meteorol. Soc.* **72**, 2–20.

Rudels, B.: 1989, 'The Formation of Polar Surface Water, the Ice Export and the Exchanges through Fram Strait', *Progr. Oceanogr.* **22**, 205–248.

Rudels, B., Jones, E. P., Anderson, L. G., and Kattner, G.: 1994, 'On the Intermediate Depth Waters of the Arctic Ocean', in Johannessen et al. (eds.), *The Polar Oceans and their Role in Shaping the Global Environment*, American Geophysical Union, 85, Washington D.C., pp. 33–46.

Savina, S. S.: 1987, 'Krupnomashtabnye Atmosfernoi Protsessy na Severnorm Polusharii i Klimaticheskie Ekstremy na Evropeisk"i Territorii S.S.R. (Large-Scale Atmospheric Processes in the Northern Hemisphere and Climatic Extremes in the European Part of the U.S.S.R.)', *Data Meteorol. Studies* **13**, Inst. of Geography, Academy of Sciences, Moscow, p. 180.

Semiletov, I. P., Savelieva, N. I., Pipko, I. I., Pugach, S. P., and Weller, G. E.: 1999, 'The Dispersion of Siberian River Flows into Coastal Waters: Hydrological and Hydrochemical Aspects', in E. L. Lewis (ed.), *The Freshwater Budget of the Arctic Ocean*, NATO Arctic Workshop, Kluwer Academic Press, Dordrecht, in press.

Serreze, M. C., Barry, R. G., and Walsh, J. E.: 1995a, 'Atmospheric Water Vapor Characteristics at 70° N', *J. Climate* **8**, 719–731.

Serreze, M. C., Maslanik, J. A., Key, J. R., and Kokaly, R. F.: 1995b, 'Diagnosis of the Record Minimum in Arctic Sea Ice Area during 1990 and Associated Snow Cover Extremes', *Geophys. Res. Lett.* **22**, 2183–2186.

Serreze, M. C., Carse, F., Barry, R. G., and Rogers, J. C.: 1997, 'Icelandic Low Cyclone Activity: Climatological Features, Linkages with the NAO, and Relationships with Recent Changes in the Northern Hemisphere Circulation', *J. Climate* **10**, 453–464.

Shaver, G. R., Billings, W. D., Chapin, F. S. III, Gibbin, A. E., Nadelhoffer, K. J., Oechel, W. C., and Rastetter, E. B.: 1992, 'Global Changes and the Carbon Balance of Arctic Ecosystems', *Bio Sci.* **61**, 415–435.

Smith, D. M.: 1998, 'Recent Increase in the Length of the Melt Season of Perennial Arctic Sea Ice', *Geophys. Res. Lett.* **25**, 655–658.

Sogaard, H., Friberg, T., Hansen, B. U., Nordstrom, C., and Christansen, T. R.: 2000, 'Carbon Dioxide Exchange of a High Arctic Ecosystem Scaled from Canopy to Landscape Level Using Footprint Modelling and Remote Sensing', *Global Biogeochem. Cycles*, in press.

Steele, M. and Boyd, T.: 1998, 'Retreat of the Cold Halocline Layer in the Arctic Ocean', *J. Geophys. Res.* **103**, 10,419–10,435.

Stocks, B. J.: 1991, 'The Extent and Impact of Forest Fires in Northern Circumpolar Countries', in Levine, J. L. (ed.), *Global Biomass Burning, Atmospheric, Climatic and Biospheric Implications*, Massachusetts Institute of Technology Press, Cambridge, MA, pp. 197–202.

Swift, J. H., Jones, E. P., Aagaard, K., Carmack, E. C., Hingston, M., Macdonald, R. W., Mclaughlin, F. A., and Perkins, R. G.: 1998, 'Waters of the Makarov and Canada Basins', *Deep. Sea Res. Part II* **44**, 1503–1529.

Thompson, D. W. J. and Wallace, J. M.: 1998, 'The Arctic Oscillation Signature in the Wintertime Geopotential Height and Temperature Fields', *Geophys. Res. Lett.* **25**, 1297–1300.

Thompson, D. W. J. and Wallace, J. M.: 2000, 'Annular Modes in the Extratropical Circulation Part I. Month-to-Month Variability', *J. Climate* **15**, 1000–1016.

Vinje, T. and Colony, R.: 1999, 'Barents Sea Ice Edge Variations over the Past 400 Years', World Climate Research Arctic Climate System Study (ACSYS), in *Proceedings of the Workshop on Sea Ice Charts of the Arctic, Seattle, WA, 5–7 August 1998*, WMO/TD No. 949, IAPO Publication No. 3, pp. 4–6.

Vinnikov, K. and Robock, A.: 1998, 'The Role of Natural Variability in Observed Global and Regional Climatic Trends', in *Preprints, Ninth Symposium on Global Change, Phoenix, Arizona*, American Meteorological Society, Boston, MA, pp. 211–214.

Wallace, J. M., Zhang, Y., and Bajuk, L.: 1996, 'Interpretation of Interdecadal Trends in Northern Hemisphere Surface Air Temperature', *J. Climate* **9**, 249–259.

Walsh, J. E., Chapman, W. L., and Shy, T. L.: 1996, 'Recent Decrease of Sea Level Pressure in the Central Arctic', *J. Climate* **9**, 480–486.

Walsh, J. E., Kattsov, V., Portis, D., and Meleshko, V.: 1998, 'Arctic Precipitation and Evaporation: Model Results and Observational Estimates', *J. Climate* **11**, 72–87.

Walsh, J. E., Vinnikov, K., and Chapman, W. L.: 1999, 'On the Use of Historical Sea Ice Charts in Assessments of Century-Scale Climate Variations', in World Climate Research Arctic Climate System Study (ACSYS), in *Proceedings of the Workshop on Sea Ice Charts of the Arctic, Seattle, WA, 5–7 August 1998*, WMO/TD No. 949, IAPO Publication No. 3, pp. 1–3.

Walsh, J. E., Zhou, X., Portis, D., and Serreze, M. C.: 1994, 'Atmospheric Contribution to Hydrologic Variations in the Arctic', *Atmos.-Ocean* **34**, 733–755.

Wang, B. and Allard, M.: 1995, 'Recent Climatic Trend and Thermal Response of Permafrost at Salluit, Northern Quebec, Canada', *Permafrost Periglacial Proc.* **6**, 221–234.

Warren, S. G., Hahn, C. J., London, J., Chervin, R. M., and Jenne, R.: 1988, 'Global Distribution of Cloud Cover and Cloud Type Amounts over the Ocean', *NCAR Tech. Note TN-317+STR*, Boulder, CO, p. 41.

Warrick, R. A., Le Provost, C., Meier, M. F., Oerlemans, J., and Woodworth, P. L.: 1996, 'Changes in Sea Level', in Houghton, J. T., Meira Filho, L. G., Callander, B. A., Harris, N., Kattenberg, A., and Maskell, K. (eds.), *Climate Change 1995, The Science of Climate Change, Contribution of Working Group I to the Second Assessment of the Intergovernmental Panel on Climate Change* (Ch. 7), Cambridge University Press, Cambridge, U.K., pp. 363–405.

Whalen, S. C. and Reeburgh, W. S.: 1992, 'Interannual Variations in Tundra Methane Emissions: A 4-Year Time Series at Fixed Sites', *Global Biochem. Cycles* **6**, 139–159.

Woo, M., Heron, R., Marsh, P., and Steer, P.: 1983, 'Comparison of Weather Station Snowfall with Winter Snow Accumulation in High Arctic Basins', *Atmos.-Ocean* **21**, 312–325.

Zakharov, V. F.: 1997, 'Sea Ice in the Climate System', *World Climate Research Programme/Arctic Climate System Study*, WMO/TD-No.782, Geneva, p. 80. Zhang, T. and Osterkamp, T. E.: 1993, 'Changing Climate and Permafrost Temperatures in the Alaskan Arctic', in *Proceedings of the Sixth International Conference on Permafrost, Beijing, China, Vol. 1,* South China University of Technology Press, pp. 783–788.

Zimov, S. A., Daviodov, S. P., Voropaev, Y. V., Prosiannikov, S. F., Semiletov, I. P., Chapin, M. C., and Chapin, F. S. III: 1996, 'Siberian CO_2 Efflux in Winter as a CO_2 Sources and Cause of Seasonality in Atmospheric CO_2', *Clim. Change* **33**, 111–120.

Zimov, S. A., Voropaev, Y. V., Semiletov, I. P., Daviodov, S. P., Prosiannikov, S. F., Chapin, F. S. III, Chapin, M. C., Trumbore, S., and Tyler, S.: 1997, 'North Siberian Lakes: a Methane Source Fueled by Pleistocene Carbon', *Science* **277**, 800–802.

Zimov, S. A., Zimova, G. M., Daviodov, S. P., Daviodova, A. I., Voropaeva, Y. V., Vorpaeva, Z. V., Prosiannikov, S. F., Prosiannikova, O. V., Semiletova, I. V., and Semiletov, I. P.: 1993, 'Winter Biotic Activity and Production of CO_2 in Siberian Soils: a Factor in the Greenhouse Effect', *J. Geophys. Res.* **98** (D3), 5017–5023.

(Received 21 April 1998; in revised form 14 June 1999)