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Glaciers in the High Arctic and Recent Environmental Change

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Phil. Trans. R. Soc. Lond. A 1995 **352**, 321-334
doi: 10.1098/rsta.1995.0073

Glaciers in the High Arctic and recent environmental change

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High Arctic climate change over the last few hundred years includes the relatively cool Little Ice Age (LIA), followed by warming over the last hundred years or so. Meteorological data from the Eurasian High Arctic (Svalbard, Franz Josef Land, Severnaya Zemlya) and Canadian High Arctic islands are scarce before the mid-20th century, but longer records from Svalbard and Greenland show warming from about 1910–1920. Logs of Royal Navy ships in the Canadian Northwest Passage in the 1850s indicate temperatures cooler by 1–2.5 °C during the LIA. Other evidence of recent trends in High Arctic temperatures and precipitation is derived from ice cores, which show cooler temperatures (by 2–3 °C) for several hundred years before 1900, with high interdecadal variability. The proportion of melt layers in ice cores has also risen over the last 70–130 years, indicating warming. There is widespread geological evidence of glacier retreat in the High Arctic since about the turn of the century linked to the end of the LIA. An exception is the rapid advance of some surge-type ice masses. Mass balance measurements on ice caps in Arctic Canada, Svalbard and Severnaya Zemlya since 1950 show either negative or near-zero net balances, suggesting glacier response to recent climate warming. Glacier-climate links are modelled using an energy balance approach to predict glacier response to possible future climate warming, and cooler LIA temperatures. For Spitsbergen glaciers, a negative shift in mass balance of about 0.5 m a^{-1} is predicted for a 1 °C warming. A cooling of about 0.6 °C, or a 23% precipitation increase, would produce an approximately zero net mass balance. A 'greenhouse-induced' warming of 1 °C in the High Arctic is predicted to produce a global sea-level rise of 0.063 mm a^{-1} from ice cap melting.

1. Introduction

Recent general circulation model (GCM) simulations of climatic response to increasing proportions of 'greenhouse gases' in the atmosphere have predicted that Arctic regions will experience enhanced warming relative to lower latitudes, particularly in winter (Cattle & Thomson 1993; Cattle & Crossley, this volume). In addition to these predictions of future, anthropogenically-induced climate change, there is evidence that the Arctic climate has fluctuated over the past few hundred years, presumably in response to 'natural' forcing factors within the climate system (Bradley & Jones 1992). These past fluctuations are exemplified by the warming in many Arctic regions associated with the ending of a period of variable but generally colder conditions, known as the 'Little Ice Age' (LIA), at about the turn of the century (Grove 1988).

Phil. Trans. R. Soc. Lond. A (1995) **352**, 321–334
Printed in Great Britain

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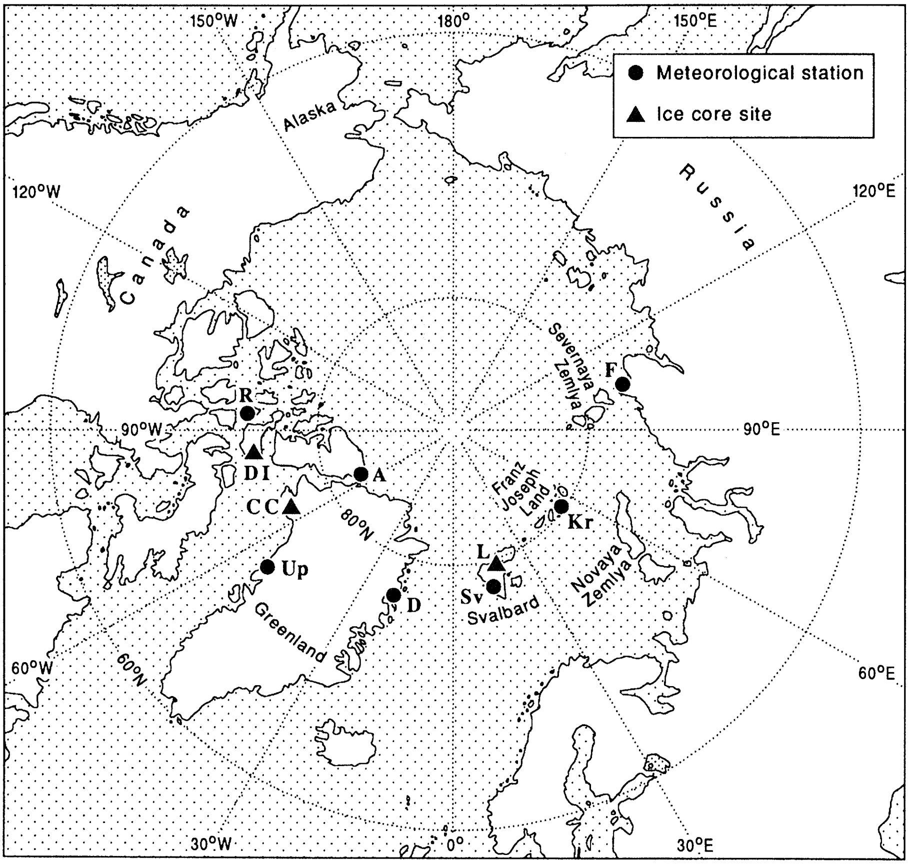

Figure 1. The location of glaciers, ice caps and ice sheets within the High Arctic. The meteorological stations and ice core sites discussed in the text are also shown: (DI) is Devon Island, (CC) Camp Century, (L) Lomonosovfonna, (R) Resolute, (A) Alert, (Up) Upernavik, (D) Danmarkshavn, (Sv) Svalbard Lufthavn, (Kr) Krenk and (F) Fedorova.

The objective of this paper is to examine how ice masses in the High Arctic have been affected by climatic fluctuations in the recent past, and how they may respond to possible future shifts in the Earth's environmental system. Recent climate change is taken here as the past few hundred years, and the next 50–200 years. The latitudinal region forming the High Arctic is defined as that area above 75° N, encompassing the northern Greenland Ice Sheet, the heavily ice covered High Canadian Arctic islands, and the Eurasian Arctic archipelagoes of Svalbard, Russian Franz Josef Land, Severnaya Zemlya and Novaya Zemlya (figure 1). The small De Long archipelago at 76–77° N and 148–158° E completes the circumpolar distribution of ice-covered High Arctic islands. The total area covered by High Arctic ice masses is about 200 000 km^2, or about 37% of the Earth's glaciers and ice caps, excluding the Greenland and Antarctic ice sheets at 1.7 and 12 million km^2, respectively (Meier 1984; Warrick & Oerlemans 1990; Oerlemans 1993).

The paper sets out the observational and proxy records of recent climatic change in the High Arctic using meteorological data and ice cores. Recent fluctuations in Arctic ice mass extent are then examined. Glacier-climate links are investigated through studies involving the measurement and energy balance modelling of glacier mass balance. Finally, the links between High Arctic ice masses and climate are discussed in the context of global sea-level rise.

2. Meteorological records of recent climate change in the High Arctic

(*a*) *Modern meteorological stations*

Meteorological data from the High Arctic before World War II are available from only a limited number of stations, reflecting the isolation of this remote, largely ice covered environment. Since the war, observations have been collected from most parts of the High Arctic, although accessible records from the Russian north are fragmentary.

Time series data for several stations around the High Arctic are shown in figure 2. The only available records from the 1930s and earlier are from Upernavik in West Greenland, Isfjord Radio (now relocated to Svalbard Lufthavn) in Spitsbergen, and Fedorova, immediately south of the heavily ice-covered Russian Severnaya Zemlya archipelago (figure 1). The most marked temperature shifts are in the two longest records. Temperatures from 1875 at Upernavik show a warming of 2 °C in mean annual temperature between the last quarter of the 19th century and the 20th, with a particularly rapid rise of about 3.5 °C if the ten years around 1920 are taken alone. The mean annual temperature in Svalbard also rose abruptly by 4–5 °C between 1912 and 1920. However, mean July temperatures at these two stations show a much less marked change over the same periods, of about 20–40% of the mean annual change (figure 2). This is significant for glacier mass balance, because summer temperatures are linked closely to glacier surface ablation, whereas a relative warming between about October and May does not lead to enhanced melting. Even so, these relatively large shifts to warmer conditions are inferred to mark the end of the LIA.

The remaining High Arctic stations in figure 2, with meteorological records of less than 50 years, show high interannual variability, but also some more consistent decadal changes. For example, the interval from 1932 to the mid-1950s was relatively warm at Fedorova, Russia, and the 1960s was a relatively cool period in Svalbard. There was a period of relative summer cooling from about 1964 to 1977 at Resolute, Canada (Bradley & England 1978), followed by warmer temperatures. The period 1920–1950 was particularly warm in West Greenland, followed by cooling of about 1.5 °C in mean annual temperature.

Two concluding points should be made on the basis of the meteorological data in figure 2. First, the records are very noisy in terms of interannual and interdecadal variability. Secondly, the records from some stations show significantly warmer and also cooler intervals during the 20th century, and it is against this background of continuing climatic variability that possible future anthropogenically-induced climate change should be set.

(*b*) *Meteorological records from the mid–19th century*

Much of the High Arctic was still being explored and mapped during the 19th century. However, some fragmentary meteorological records do exist from this period, mainly in ships' logbooks. The era of the search for the Canadian Northwest Passage (approximately 74° N), and Sir John Franklin, saw over 25 wintering ships in this area between about 1819 and 1859. Each ship was frozen into the stable winter sea-ice

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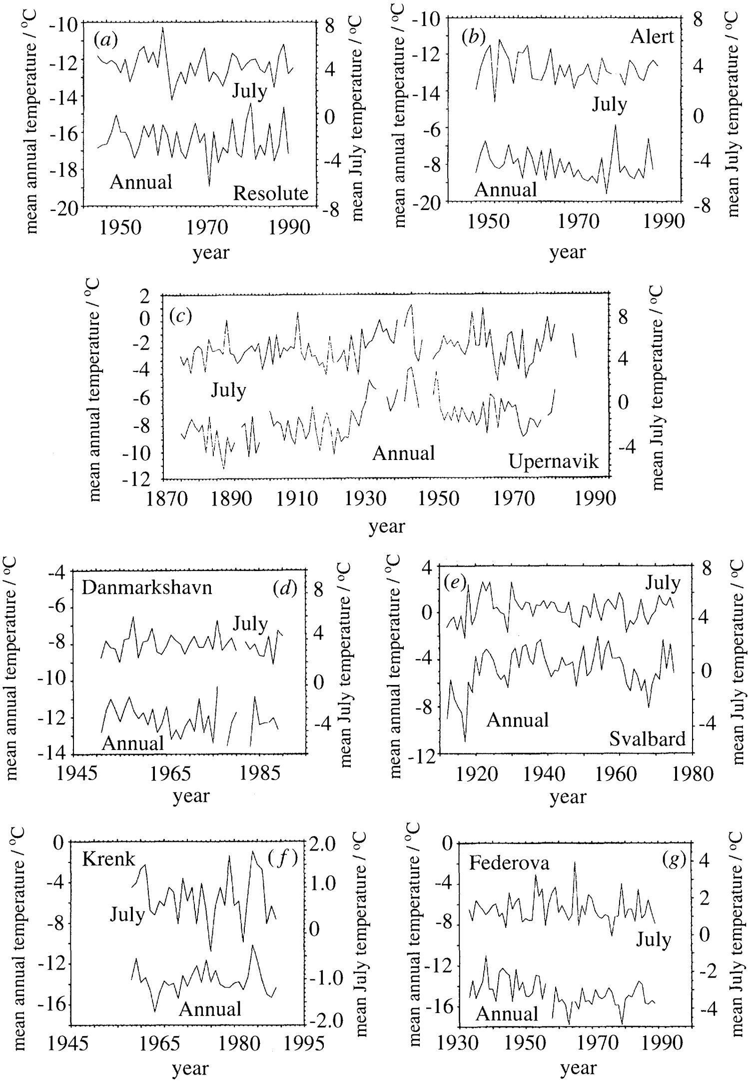

Figure 2. Meteorological records of fluctuations in mean annual and mean July temperature at stations in the Canadian, Greenland and Eurasian High Arctic sectors (located in figure 1). (*a*) Resolute, Cornwallis Island, Canada (74°43′ N, 95°59′ W). (*b*) Alert, Ellesmere Island, Canada (82°30′ N, 62°20′ W). (*c*) Upernavik, West Greenland (72°47′ N, 56°10′ W). (*d*) Danmarkshavn, East Greenland (76°46′ N, 18°40′ W). (*e*) Longyearbyen, Svalbard (78°04′ N, 13°38′ E). (*f*) Krenk Station, Hayes Island on Franz Josef Land, Russia (80°37′ N, 58°03′ E). (*g*) Fedorova, northern Taymyr Peninsula, Russia (77°43′ N, 104°17′ E).

cover, providing a fixed site at sea level for systematic meteorological observations within the LIA (Grove 1988).

The logs of these Royal Navy ships are being analysed (Dowdeswell & Barr 1995). Records of temperature, pressure, wind speed and direction, and sometimes precipitation were logged, with measurements a number of times each day. Instruments were housed away from the ship, elevated from the ground and shaded. Reproducibility has been investigated by comparisons of temperature and pressure data recorded in-

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dependently by the ships Resolute and Intrepid, locked together in sea ice in 1852–53. Correlation coefficients between the two sets of observations are better than 0.9–0.95.

July temperatures have so far been extracted from the logs of ten ships trapped in shorefast sea ice at locations within 1.5° of latitude 75° N between 1849 and 1859. Mean July maximum temperatures are an average of 4.8±1.4 °C for the ten vessels. July values are presented, since summer temperatures are a first order control on glacier surface melting and on net mass balance. The positions of these ships lie around the modern weather station of Resolute (Bradley & England 1978). Comparing the mean July maximum values for the mid-19th century with modern records shows that only once from 1948–1976 was the observed value below the mid-19th century mean. Modern values are a mean of 7.5±1.3 °C for 1948–63 and 6.0±1.6 °C for 1964–1976. Direct observations from the Canadian High Arctic in the mid-19th century therefore support the evidence of proxy climate datasets, that this was a colder period relative to most of the 20th century, with summers cooler by about 1–2.5 °C.

3. Recent climate change from High Arctic ice-core records

Deep ice cores have been obtained from a number of High Arctic ice caps (figure 3). Where little or no surface melting takes place to complicate core stratigraphy and chemistry, a temporal resolution of ±1–2 years is possible over the past few centuries (Bradley 1985). However, in the Eurasian High Arctic, where mean annual temperatures are significantly higher, melting and refreezing effects dominate and chronology is more difficult to establish (Dowdeswell *et al.* 1990; Tarussov 1992).

Several parameters have been used to investigate climate change in High Arctic ice cores, including: (i) oxygen isotope ratios, which are related to temperature at the time of snow condensation and several complicating factors (Bradley 1985); and (ii) the number and thickness of refrozen melt layers, which are proportional to ice surface melting and therefore provide an index of summer warmth (Koerner 1977). The results of oxygen isotope analyses of ice cores from the High Arctic vary in detail (figure 3), but some broad similarities are present in isotopic records from Devon Island, Canada (Paterson *et al.* 1977), through Camp Century in North Greenland (Johnsen *et al.* 1970), to Lomonosovfonna in Svalbard (Gordiyenko *et al.* 1981) and the Vavilov Ice Dome, Russian Severnaya Zemlya (Kotlyakov *et al.* 1990). The previous two to three centuries have markedly more negative oxygen isotopic ratios than the 20th century (figure 3). The 16th century is somewhat warmer than the 1600–1900 period, but isotopic values do not generally approach those of the relatively warmer 20th century (figure 3). Changes towards less negative ratios over the last 100–150 years are about 1.5 ppt on the Canadian Arctic ice caps, representing a temperature rise of about 2.5 °C over LIA conditions. Core chronologies are most uncertain for the Eurasian High Arctic, due to the intense nature of surface melting.

The recent records of melt layers in Arctic ice cores also show similarities over space, and with the isotopes. On Devon Island the period since about 1860 has shown a rise in the number and thickness of melt layers relative to the preceding 300 years, with summer warming intensifying since the 1920s (Koerner & Paterson 1974; Koerner 1977). Ice cap mass balance is now negative, but is predicted to have been close to zero for the earlier, cooler period (Koerner 1977). The very low frequency of melt layers about 150 years ago indicates that this was the interval of coldest summers

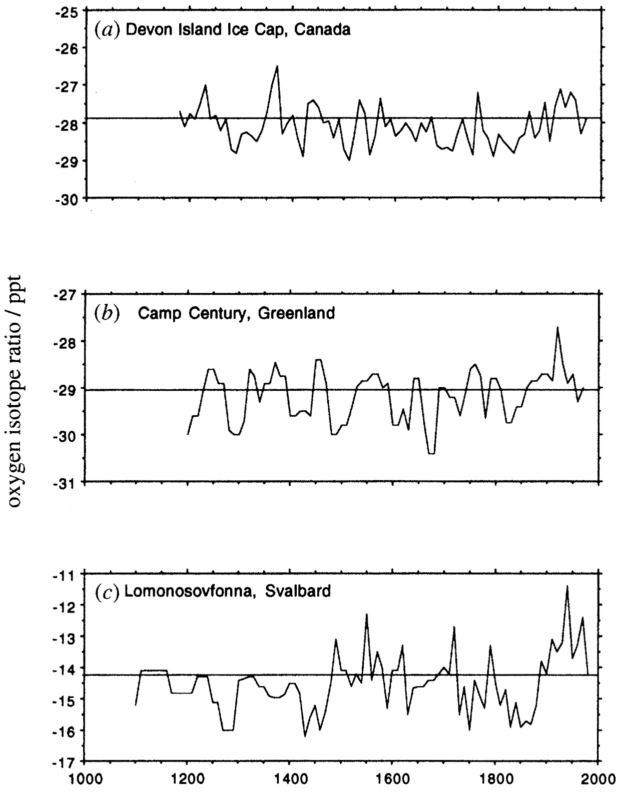

Figure 3. Oxygen isotope ratios over the last thousand years recorded in ice cores from three ice masses in the High Arctic. Horizontal lines in each diagram represent modern isotope values. The core sites are located in figure 1. (*a*) Devon Island Ice Cap, Canada (from Paterson *et al.* 1977). (*b*) Camp Century, North Greenland (from Johnsen *et al.* 1970). (*c*) Lomonosovfonna, Svalbard (from Gordiyenko *et al.* 1983).

over the whole Holocene (Koerner & Fisher 1990). Analysis of melt features in the Dye 3 core, Greenland, suggests a similar cool period from 1530–1860 (Herron *et al.* 1981). In Svalbard ice cores, the interval interpreted to represent from about 1550 to 1920 contains up to 30–40% less refrozen ice layers than that since 1920 (Tarussov 1992). The first half of the 16th century is intermediate between these two periods. The melt layer signal from the large ice caps in the Russian archipelago of Severnaya Zemlya indicates a warming trend from about 120–140 years ago, earlier than in Svalbard (Kotlyakov *et al.* 1989; Tarussov 1992).

The isotopic and stratigraphic evidence from High Arctic ice cores indicates that the cold LIA began to ameliorate from about 1860–1880 in the Canadian Arctic, Greenland and the eastern Eurasian Arctic, but that warming took place somewhat later in Svalbard. Direct meteorological observations from western Svalbard support the ice core evidence on the timing of warming, with temperatures increasing dramatically between 1912 and 1920 (figure 2*e*). However, despite some broad agreement, the detailed isotopic and stratigraphic records from individual High Arctic ice cores show high interannual to interdecadal variability.

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4. Evidence of recent glacier fluctuations in the High Arctic

(a) The retreat of glacier and ice cap margins

Fluctuations in the position of glacier termini can reflect changes in climate, through the effects of shifts in temperature and precipitation on glacier mass balance. However, the links between climate and glacier fluctuations are not simple and glacier dynamic factors must also be considered. Ice masses in most areas of the High Arctic have been retreating for much of the 20th century, whether their margins end on land or in marine waters. On Svalbard, aerial photographs acquired at intervals since the 1930s show that many glaciers are in retreat from clearly defined terminal moraine systems, with the exception of those that have surged. Chronological control for the last few hundred years is usually provided by radiocarbon and lichenometric dating methods, and Werner (1993) has used calibrated lichen growth curves to show that glaciers retreated from prominent moraine systems in north and north-western Spitsbergen from about the turn of the century. In the Canadian High Arctic, a number of glaciers have been retreating over the last century or so. However, the terminus positions of many glaciers and ice caps in Arctic Canada have not changed significantly over the past few decades, although shrinkage and thinning were measured on some especially during the period 1950–1970 (R. M. Koerner, quoted in Meier 1984).

Observational evidence from the 1930s onwards also suggests that ice masses have been retreating throughout most of the Russian High Arctic (37–158° E), and scattered local accounts are available earlier. Comparison of ice margins in Franz Josef Land between aerial photographs obtained in the 1950s and recent satellite imagery suggests a general retreat, in some cases of 2–3 km. In Novaya Zemlya, measurements between the 1930s and 1950s indicated an average retreat of 2–3% (Chizhov & Koryakin 1962). On Severnaya Zemlya, Pioneer Glacier was observed to shrink by 27 km^2 or 11% of its area between the 1930s and 1978, and Govorukha *et al.* (1987) report the loss of about 500 km^2 of ice-covered area throughout the islands between 1931–84. Significant shrinkage of the 80 km^2 ice cover on the de Long archipelago in the Asian High Arctic has also taken place over the last 30–40 years (Verkulich *et al.* 1992). Summarizing these observations, Koryakin (1986) states that during the 20th century Novaya Zemlya has lost 830 km^2 or about 3% of its glacierized area, Franz Josef Land about 700 km^2 or 5%, but suggests, in conflict with Govorukha *et al.* (1987), that ice on Severnaya Zemlya has undergone only minor retreat in limited areas. Together, this amounts to an estimated loss of total ice volume of between 1 200 and 1 600 km^3 this century (Koryakin 1986).

(b) Recent advances of glaciers and ice caps

There are two main exceptions to the broad picture of recent glacier retreat in the High Arctic. Neither is related directly to climate change. First, ice masses in several areas are known to undergo periodic rapid advances or surges, unrelated to climate, punctuating longer periods of stagnation and slow terminus retreat (Meier & Post 1969). Surge-type glaciers are concentrated in some areas, whereas in others the phenomenon does not seem to take place. Within the High Arctic the greatest concentration of known surge-type glaciers is in Svalbard (Dowdeswell *et al.* 1991). The looped moraines indicative of previous surge activity have also been reported from East Greenland (Weidick 1988). Rapid glacier advances resulting from surges are only rarely documented from the Russian and Canadian Arctic archipelagoes (Hattersley-Smith 1969).

Secondly, the marine termini of ice masses are highly sensitive to water depth. This is demonstrated by a strong dependence of the rate of mass loss through iceberg production on water depth (Brown *et al.* 1982). For example, the very rapid retreat of ice through the fjords of Glacier Bay, Alaska, is linked to deep water inland of shallow sills, with glacier margins undergoing very rapid retreat after a small initial change destabilized their position on shallow pinning points (Brown *et al.* 1982). Recent fluctuations of a series of fast-flowing tidewater glaciers in southwest Greenland have been variable. Some of these glacier termini have retreated, whereas others have advanced or remained stable. Warren (1991) relates these variable marginal fluctuations to the influence of differing bathymetry on the dynamic behaviour of individual glacier margins, rather than to climatic variability.

(*c*) *Glacier response time to climate change*

A time lag between climate change and glacier terminus response is present because mass balance is perturbed throughout glacier length, but is transferred down-glacier at finite velocities over a range of distances. The effect of a climatic shift therefore arrives at a glacier margin over a period, and the terminus position is then a weighted mean of past climate changes over the time interval (T_{m}) beyond which there is no memory of former climate (Johannesson *et al.* 1989). T_{m} is the time constant in an exponential, asymptotic approach to a new steady state after a given shift in climate. Johannesson *et al.* (1989) propose that

$$T_{\mathrm{m}} = h/(-b_{\mathrm{t}}), \tag{1}$$

where h approximates to maximum glacier thickness, and b_{t} is the mass balance at the terminus, which is a negative value. This yields timescales for adjustment to changing mass balance on the order of 10^2 years for many Arctic valley and outlet glaciers, which provides an approximation to the lag time between climatic variations and glacier response. Clearly, the larger ice caps in the High Arctic, where ice may be over 500 m thick and mass loss at the margins is relatively small, will have a longer lag time of a few hundred years.

5. Measurement and modelling of glacier-climate links

(*a*) *The mass balance of High Arctic glaciers*

The relationship between inputs of mass to High Arctic glaciers, as snow or refrozen meltwater, and mass loss, in the form of meltwater runoff and iceberg production (for tidewater glaciers), is known as glacier mass balance. If, over a balance year, inputs exceed losses, then an ice mass has a positive balance, and *vice versa*. However, long time series of glacier mass balance observations are scarce and losses by iceberg calving are very difficult to quantify, restricting mass balance data largely to ice masses ending on land.

Time series mass balance data are available from several High Arctic ice masses (figure 4). In each region, mean mass balance since the 1950s has been negative, implying a net loss of mass, albeit small in the cases of the Vavilov Ice Dome in Severnaya Zemlya and the Devon Island Ice Cap, Canada. Consistently negative glacier mass balance data have also been reported from Franz Josef Land ice masses

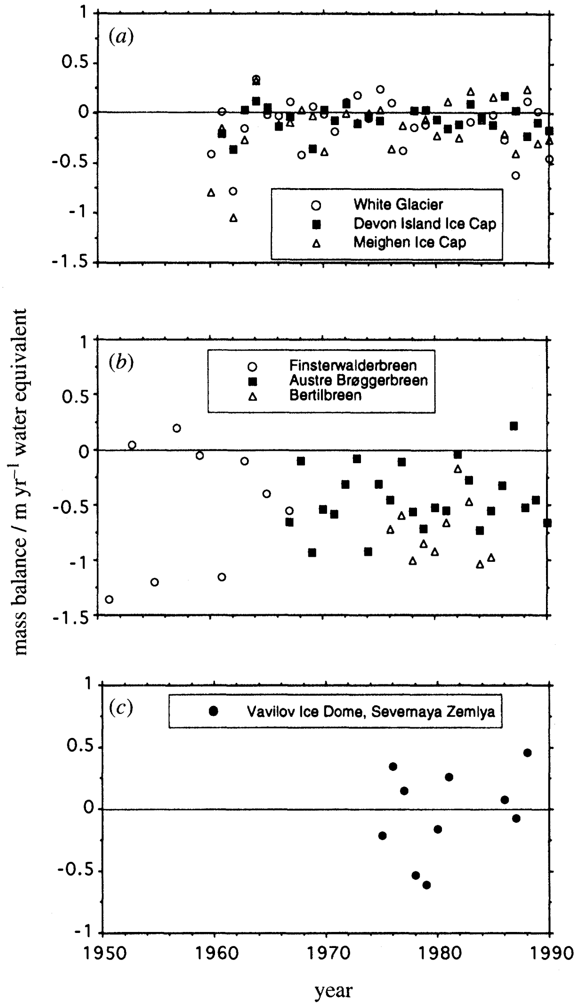

Figure 4. Trends in glacier mass balance from the High Arctic over the last 45 years. The mean and standard deviation of data for each glacier is given in water equivalent. (*a*) Meighen Ice Cap (-0.13 ± 0.29 m a^{-1}), Devon Island Ice Cap (-0.06 ± 0.13 m a^{-1}), and White Glacier on Axel Heiberg Island (-0.10 ± 0.26 m a^{-1}), Canadian High Arctic (from Koerner (1977) and World Glacier Monitoring Service). (*b*) Finsterwalderbreen (-0.51 ± 0.59 m a^{-1}), Austre Brøggerbreen (-0.42 ± 0.30 m a^{-1}) and Bertilbreen (-0.74 ± 0.27 m a^{-1}), Svalbard (Hagen & Liestøl 1990). (*c*) Vavilov Ice Dome, Severnaya Zemlya (-0.03 ± 0.36 m a^{-1}) (Barkov *et al.* 1992).

(Grosswald & Krenke 1962). All the mass balance datasets have a relatively high interannual variability, and it is only in Svalbard that the standard deviations fall entirely below the line representing mass balance equilibrium (figure 4). Even so, it can be concluded that, during the last 30–40 years, glaciers from most areas of the High Arctic have shown no sign of building up; a finding that is in line with meteorological and ice core evidence indicating climate warming over the last century or so (figures 2 and 3).

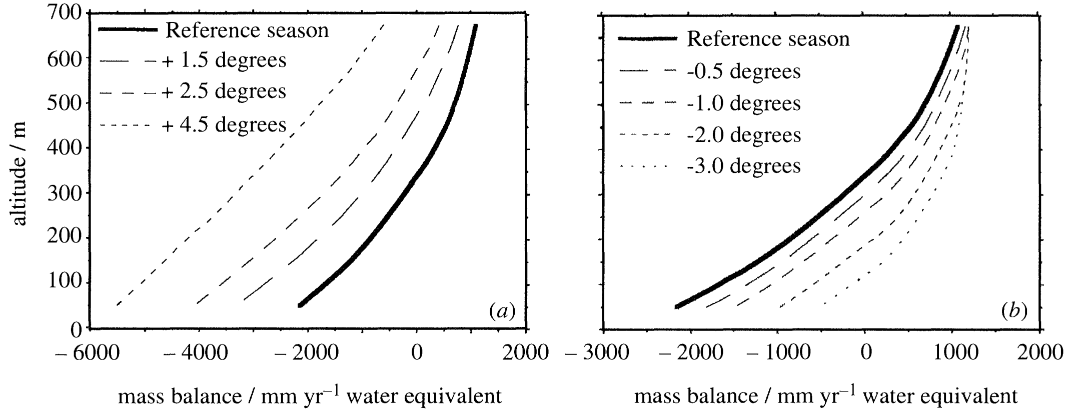

Figure 5. Energy balance modelling of the mass balance of a northwest Spitsbergen glacier in response to recent climate change (after Fleming *et al.* 1995). Modelled mass balance with altitude with: (*a*) increases in annual mean temperature of 1.5, 2.5 and 4.5 °C; (*b*) an envelope of cooler temperatures. Calculations using a modern reference climate are shown for comparison.

(*b*) *Energy balance models of glacier–climate interactions*

Energy balance modelling of the mass balance of High Arctic glaciers provides a method of assessing their sensitivity to climatic shifts. The model, which calculates the components of ice surface energy balance, takes meteorological data, the area distribution with altitude of the ice mass, and parameters defining the global radiation as input values (Oerlemans 1993). The mass balance of a glacier surface is expressed as:

$$M = \int_{\text{year}} [(1-f)\min(0, -B/L) + P]\,\mathrm{d}t, \tag{2}$$

where M is annual mass balance, f is the fraction of melt water that refreezes instead of running off, B is the energy balance of the surface, L is the latent heat of melting and P is the rate of solid precipitation. The energy balance is found from:

$$B = Q(1-a) + I_{\text{in}} + I_{\text{out}} + F_{\text{s}} + F_{\text{l}}, \tag{3}$$

where a is the surface albedo, Q is the shortwave radiation reaching the surface, I_{in} and I_{out} are the incoming and outgoing longwave radiations and F_{s} and F_{l} are the sensible and latent heat fluxes. Details of the model are in Oerlemans (1993).

Model results for several Spitsbergen glaciers, using observed meteorological parameters, yielded good predictions of measured mass balance over the last ten years (Fleming *et al.* 1995). Average net balances for 1980–1989, predicted using models tuned to the average for the decade, were −0.44 m and −0.47 m water equivalent for two north-west Spitsbergen glaciers, compared with measured averages of −0.27 m and −0.36 m. The model was then used to predict the effects of recent climate change on glacier mass balance and equilibrium line altitude. Several climate warming scenarios were input to the model (figure 5), which predicted a negative shift in net mass balance of 0.5–0.8 m a^{-1} for each degree of warming (Fleming 1992), depending on the area/elevation distribution of individual glaciers. By contrast, modelling suggested a cooling of about 0.6 °C, or a precipitation increase of around 23%, would be required to give a zero or slightly positive net mass balance (figure 5).

Table 1. *Predicted contribution of the major High Arctic ice-covered regions to global sea-level rise for a 1 °C warming and adjusted precipitation, according to equation (4). Equation and data on ice cover and precipitation are from Oerlemans (1993)*

region of the high Arctic	ice-covered area / km^2	precipitation / mm a^{-1}	sea-level rise / mm a^{-1}
Svalbard, Arctic Norway	36 600	375	0.018
Franz Josef Land, Russia	13 700	312	0.005
Novaya Zemlya, Russia	23 600	500	0.016
Severnaya Zemlya, Russia	18 300	250	0.005
Ellesmere Island, Canada	80 000	220	0.014
Axel Heiberg Island, Canada	11 700	220	0.002
Devon Island, Canada	16 200	225	0.003
total for the High Arctic	200 100	—	0.063

6. Implications for global sea-level change

The possible contribution of High Arctic ice masses to 'greenhouse-induced' global sea-level rise can also be estimated, following Oerlemans (1993). A strong relation has been found between glacier sensitivity and mean annual precipitation, which quantifies the relatively low sensitivity of High Arctic ice masses to climate change compared with glaciers in lower latitudes (Oerlemans & Fortuin 1992). Energy balance modelling experiments on twelve glaciers from varying latitudes and climatic regimes were used to predict mass balance averaged over the entire glacier (B_m). A one degree temperature increase was simulated, in which precipitation was increased in proportion to the saturation vapour pressure calculated from the mean annual air temperature over each glacier (Oerlemans 1993). The experiments gave the following relation concerning mass balance change (∂B_m):

$$\partial B_m = -0.401 - 0.514 \log P, \tag{4}$$

where P is mean annual precipitation, indicating that High Arctic ice masses, with precipitation often of only a few tens of centimetres per year, are less sensitive to climate change than those in warmer and more maritime regions (Oerlemans & Fortuin 1992; Oerlemans 1993).

Applying equation (4) to the ice masses in the High Arctic, excluding the Greenland Ice Sheet, provides an estimate of the contribution of each area to global sea-level rise for a 1 °C warming (table 1). The combined contribution of ice masses in the Canadian, Norwegian and Russian High Arctic gives a global sea-level rise of 0.063 mm a^{-1}. This total represents about 14% of the 0.46 mm a^{-1} sea-level rise predicted by Oerlemans (1993) for the contribution of glaciers and ice caps from all latitudes, excluding Greenland and Antarctica. A problem with these calculations is that values for precipitation are based on few data and, although unlikely to vary by an order of magnitude, will probably require revision as new information becomes available. The complex issue of the influence of the Greenland Ice Sheet on global sea-level is considered elsewhere (Wingham, this volume).

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7. Conclusions

Climate change in the High Arctic over the last few hundred years has included the relatively cool LIA, followed by recent warming. However, these broad climatic trends are represented in individual locations by high interannual and interdecadal variability in the climate system (figure 2). Long meteorological records from Svalbard and Greenland show warming of several degrees from the late 19th to early 20th centuries (figure 2). Proxy evidence of High Arctic climate is derived mainly from ice cores, and oxygen isotopic records show generally cooler temperatures (by 2–3 °C) for several hundred years preceding 1900 (figure 3). The proportion of melt layers in ice cores has also risen significantly over the last 70–130 years, indicating recent summer warming. Ice masses in the High Arctic (figure 1), have responded to these changes in climate in a number of ways.

(i) There is widespread geological evidence of glacier retreat in the High Arctic since the end of the LIA. An exception to this is the rapid advance of some ice masses during surges, unrelated to climate forcing.

(ii) Mass balance measurements on High Arctic ice caps since the 1950s show that net balances are either negative or near-zero (figure 4), suggesting that glaciers are responding to recent climate warming.

(iii) Glacier-climate links have been modelled using an energy balance approach to predict the response of glaciers to future and past climate change. For Spitsbergen glaciers (figure 5), a negative shift in mass balance of about 0.5 m a^{-1} is predicted for a 1 °C warming. A cooling of about 0.6 °C, or a 23% precipitation increase, would produce an approximately zero net mass balance.

(iv) A 'greenhouse-induced' warming of 1 °C affecting glaciers throughout the High Arctic is predicted to produce a global sea-level rise of 0.063 mm a^{-1} from ice cap melting (table 1). This is about 14% of the modelled sea-level contribution of all glaciers and ice caps, excluding the Antarctic and Greenland ice sheets.

In conclusion, predictions of the possible future response of High Arctic ice masses to environmental change should be set within the context of continuing climatic variability which is not forced exclusively by anthropogenically-induced changes to the atmospheric system, as exemplified by the direct and proxy records of climate change for the last few hundred years from the High Arctic (figures 2–4).

Parts of this work have been supported by the U.K. Natural Environment Research Council (Grant GR3/8507), the European Community (Grants EV5V-CT93-0299 and INTAS-1010-CT93-0006) and The Royal Society. I thank E. K. Dowdeswell for collating meteorological and mass balance data and R. M. Koerner for comments.

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