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HOLOCENE PALEOCLIMATOLOGY OF THE QUEEN ELIZABETH ISLANDS, CANADIAN HIGH ARCTIC

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A wide variety of evidence reflecting, in different ways, the changing climate of the Queen Elizabeth Islands during the Holocene is reviewed. All proxies pertain to summer conditions. Many sources of information provide weak or equivocal paleoclimatic signals, but a general pattern of events can be discerned. Temperatures in the early to mid-Holocene were highest (comparable with, or higher than, temperatures prevailing for much of this century). Although much evidence points to a mid-Holocene thermal maximum, there is also considerable evidence that conditions were warmer in the early Holocene, possibly related to orbitally-induced radiation anomalies. The apparent mid-Holocene thermal maximum may reflect lags in the response of the environment and of some proxies in recording paleoclimatic conditions, more direct indices point to warmest conditions in the early Holocene (before 7500 BP). Temperatures declined from ~3000 BP culminating in exceptionally low temperatures from 100–400 BP This may have been the coldest period of the entire Holocene, resulting in glacier advances to post-glacial maximum positions. The period since 1925 has witnessed a pronounced increase in temperature, leading to negative mass balances on glaciers and ice sheets throughout the archipelago. This period is the warmest for at least 1000 years and perhaps for several thousand years. Modern climate in the Queen Elizabeth Islands may thus be characteristic of the early to mid-Holocene and atypical of conditions which prevailed for much of the last few thousand years.

INTRODUCTION

The Queen Elizabeth Islands comprise an archipelago of about 1.3 million square kilometers, equivalent to an area the size of Fennoscandia, or of the United States east of the Mississippi/Ohio rivers, excluding Florida (Fig. 1) The islands are mountainous and glacierized in the east with large ice caps on Axel Heiberg, Ellesmere and Devon Islands The highest peaks reach to over 2500 m in the Grant Land Mountains, Ellesmere Island The western islands are generally lower in relief and currently unglacierized. The only exceptions are three small glaciers on Melville Island and a small ice cap on Meighen Island (Paterson, 1969) Glaciation levels and equilibrium line altitudes in the region have been mapped and discussed by Miller *et al* (1975) In general, glaciation levels are lowest along the northern and northwestern coasts of the region, more or less reflecting the pattern of minimum melting degree day totals (Edlund and Alt, 1989).

The paleoclimatology of the area is not well-known, with questionable interpretations of proxy records often buttressed by dubious data from adjacent regions. Here the evidence currently available from different proxies is critically reviewed and the strongest arguments are then summarized. The paper complements a similar review of Holocene paleoclimatology in the Baffin Island area (Williams and Bradley, 1985)

MODERN CLIMATE

Climatically, the area can be described as polar desert or semi-desert (Bovis and Barry, 1974) Contemporary climatic information is sparse, based largely on a handful of synoptic observing stations at coastal, low elevation sites (Resolute, Mould Bay, Isachsen [now closed] Eureka, Alert and Thule (N W Greenland)) supplemented by short-term (generally summertime) observations at research camps, co-ordinated by the Canadian Polar Continental Shelf Project, Ottawa

Recent studies of vegetation patterns in the region, and detailed analysis of all the available climatic data, indicate that climatic conditions vary across a northwest–southwest gradient, from the Arctic Ocean towards Baffin Bay (Edlund and Alt, 1989) As the Arctic Ocean is approached, summer temperatures are lower, the melt season is shorter (and starts later) and cloudiness increases These conditions are reflected in the pattern of vegetation which forms an ‘S-shaped’ pattern, with the lowest number of vascular plant species (<50) and lowest percent vascular plant cover (<10%) in a region centered on Lougheed Island (Edlund, 1983, 1986) Within this bioclimatic zone, vegetation is entirely herbaceous. By contrast, some areas further south and east have over 100 vascular plant species, over 25% vascular plant cover and the vegetation is dominated by sedges and dwarf shrubs (Edlund, 1984)

For most of the year, temperatures are well below 0°C and a strong surface temperature inversion prevails (Bilello, 1966, Maxwell, 1981) The melt season is generally less than 85 days (early June to late August) with the longest melt periods experienced in low elevation continental locations, such as Eureka (Fig 1). Mean July temperatures reflect the S-shaped bioclimatic zonation discussed earlier, being lowest in the region centered on Lougheed Island and extending along the northwestern coastal region, but higher to the south and east (Fig. 2) Highest mean July tempera-

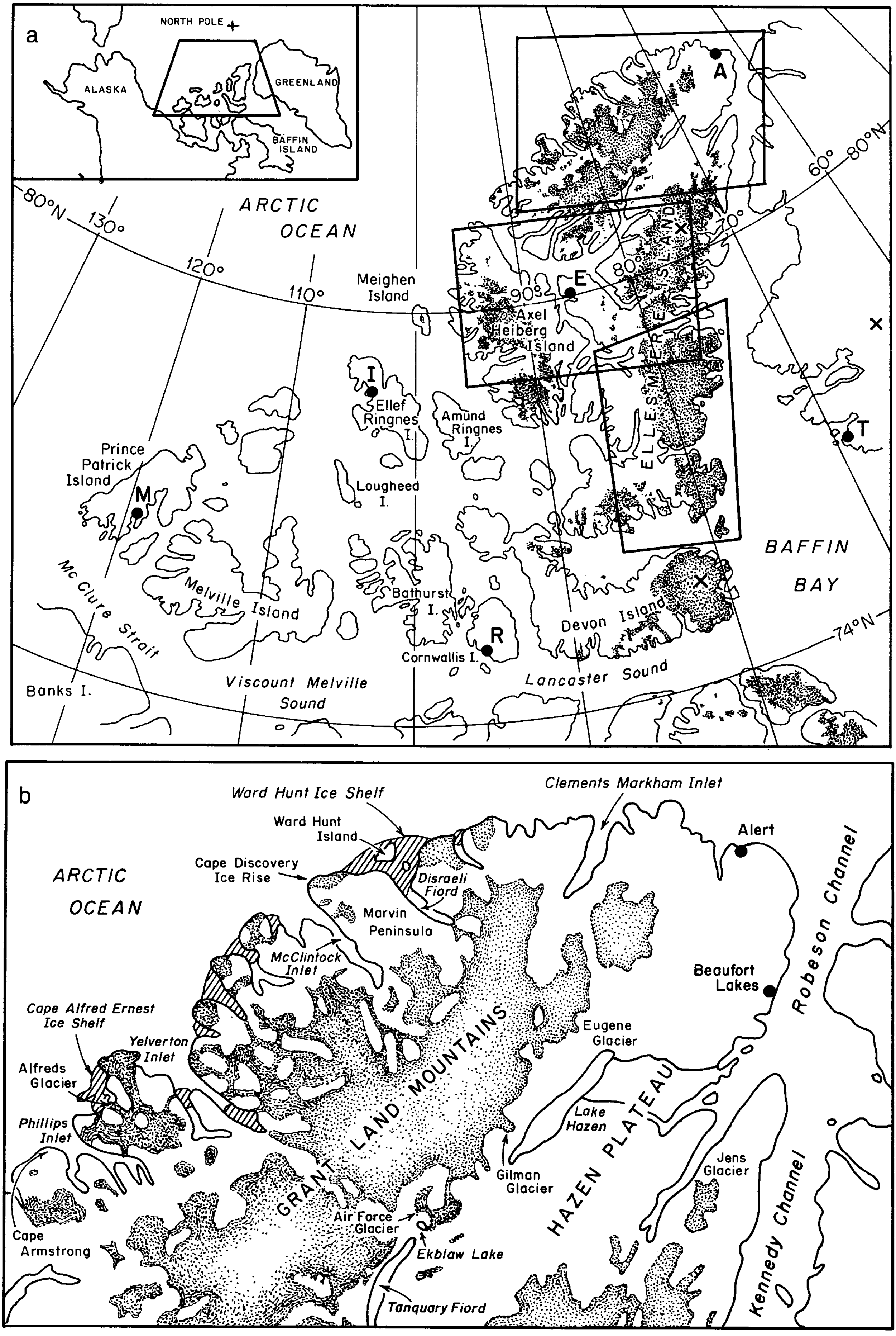

FIG. 1a,b. The Queen Elizabeth Islands, Northwest Territories, Canada and location of places mentioned in the text.

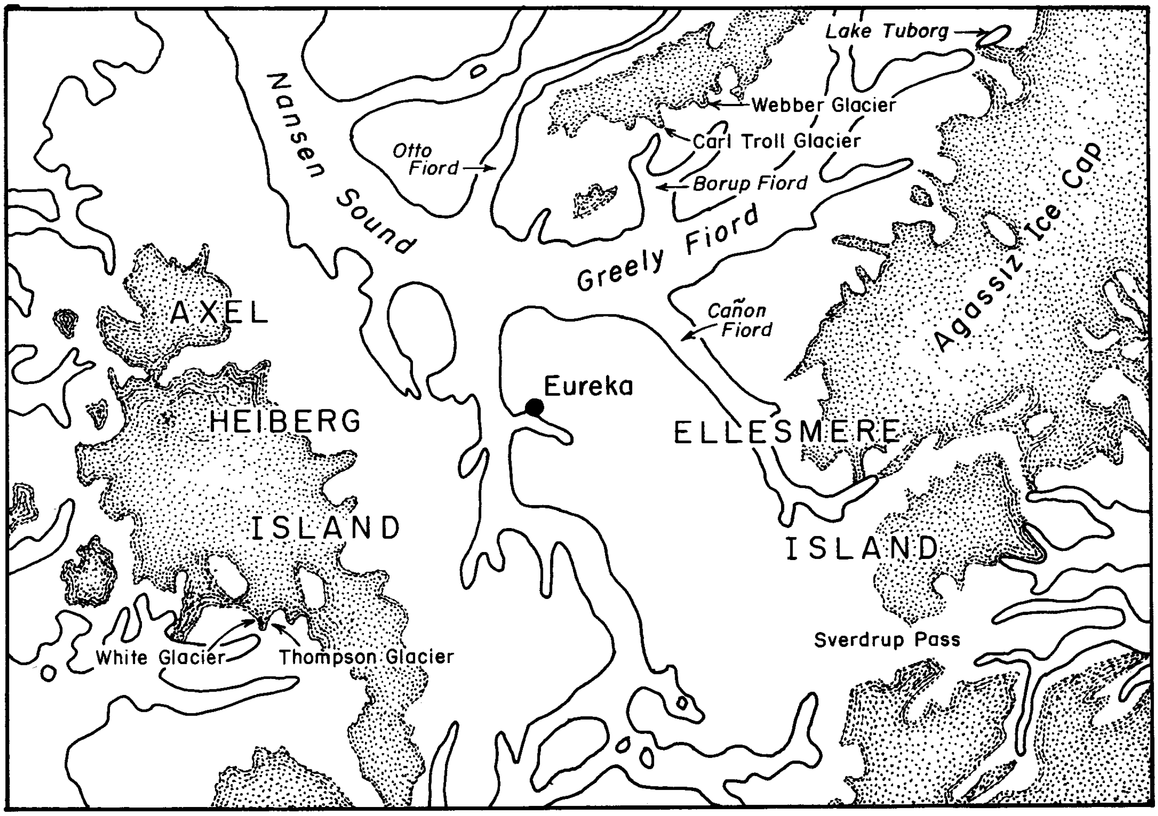

FIG 1c. Queen Elizabeth Islands, Northwest Territories, Canada and location of places mentioned in the text.

tures (>8°C) are recorded in west-central Ellesmere Island, due to the effects of continentality (Edlund and Alt, 1989)

The inter-island channels are ice-covered from late September until late May except for a few persistent polynias. Ice shelves currently exist along the northern and northwestern coastline of Ellesmere Island where mean summer temperatures (June–August) are <−2°C and persistent cloud cover is common (Edlund and Alt, 1989) Precipitation is low (<17 cm per year) over most of the archipelago (Bradley and England, 1978) but increases significantly, to over 30 cm per year in the southeastern margins, especially on the east-facing slopes of high elevation ice caps and glaciers facing Baffin Bay (Koerner, 1983) Rainfall in summer months is not uncommon, but the bulk of precipitation is in the form of snow, with seasonal maxima in both early and late winter months (particularly late August–early November) Moisture-bearing storms enter the region from the southwest (across the Beaufort Sea), from Baffin Bay or from the west and northwest (across the Arctic Ocean) (Bradley and England, 1978), Bradley and Eischeid, 1985, Bourgeois *et al.*, 1985) Most of the annual precipitation results from a relatively small number of synoptic events. At Alert, for example, half of the annual precipitation accumulates on about 20 days per year, on average (Bradley and Eischeid, 1985). Further discussion of the region's climate can be found in Wilson (1967), Maxwell (1980, 1981, 1982), and Edlund and Alt (1988, 1989).

EVIDENCE FOR PALEOCLIMATIC CONDITIONS IN THE HOLOCENE

Sources of paleoclimatic information are limited to a few, often equivocal types of record, most of which are interpreted as proxies of summer temperature Only ice cores and lake sediments provide continuous records, but these are often difficult to interpret and must be supplemented by a wide variety of other data to enable a broad-scale picture of events to be reconstructed

Ice Core Records

Agassiz Ice Cap

Three cores to bedrock have been recovered from near the crest of Agassiz Ice Cap at ~80°N on eastern Ellesmere Island (Fisher *et al*, 1983, Fisher and Koerner, 1988, Koerner *et al*, 1988) In the case of core A79, Fisher *et al* (1983) used a flow model to calculate the age of the ice at depth, assuming accumulation rates have not changed from recent averages. This assumption is supported by the matching of high conductivity peaks in the core (indicative of major volcanic eruptions) with similar peaks found in core A77 and at Camp Century, Greenland (dated independently) (Paterson and Waddington, 1984) The chronology of cores A77 and A79 is estimated to be accurate to within ± 10% back to 6500 (calendar

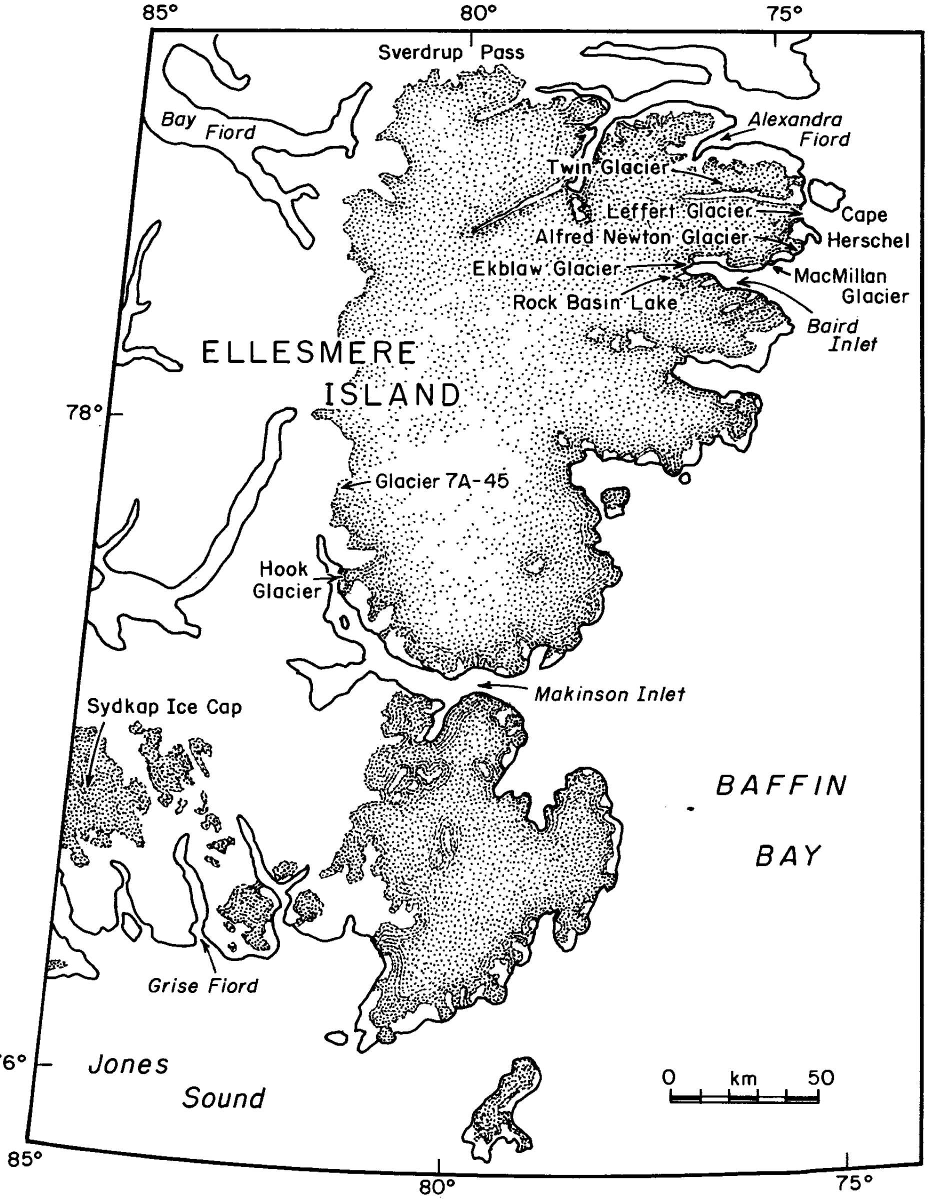

FIG 1d. Queen Elizabeth Islands, Northwest Territories, Canada and location of places mentioned in the text.

years*) BP (core A79) or ± 10% back to 7000 BP (core A77) (Fig. 3) A large increase in $\delta^{18}O$ near the bottom of each core is taken to be synchronous with similar changes in Camp Century and Dye-3 ice cores (from Greenland) which are dated around 10,500 (calendar years) BP (Hammer *et al.*, 1986; Dansgaard *et al.*, 1989).

As a result of scouring by katabatic winds around the ice core sites, the oxygen isotope profiles 'are of limited value as climatic records' (Fisher *et al.*, 1983) However, the record of the last 1500, and perhaps 3000 years, from core A77 is considered to have not been affected significantly by scouring. $\delta^{18}O$ values over the last 3000 years have declined by 1 7‰, if the relationship between $\delta^{18}O$ and mean annual temperature is meaningful (Dansgaard *et al.*, 1973), this implies a cooling of ~3°C in 3000 years. Since similar trends occur in core A79 (where scouring of mainly winter accumulation has occurred) they suggest that trends in summer temperature have been similar to trends in mean annual temperature, and during the 'Little Ice Age' (lasting from ~1650 to 1900 in this record) summer temperatures may have actually cooled more than mean annual temperatures (Fisher *et al* , 1983) Based on a reduction in the seasonal amplitude of $\delta^{18}O$, Fisher and Koerner (1988) argue that the winter sea–ice boundary was further south by 3.5° latitude during the so-called 'Little Ice Age' However, it seems as probable that greatly reduced open water conditions in summer (Alt *et al.*, 1985) contributed to the reduction in seasonal $\delta^{18}O$ amplitude.

Analysis of pollen in the Agassiz ice core A77 reveals three principal depositional periods (Bourgeois, 1986) (Fig. 4). From the beginning of the Holocene to ~7600 BP, pollen concentrations were extremely low (~14

*Dates given for ice cores are in calendar years and other dates are in radiocarbon years, unless otherwise noted.

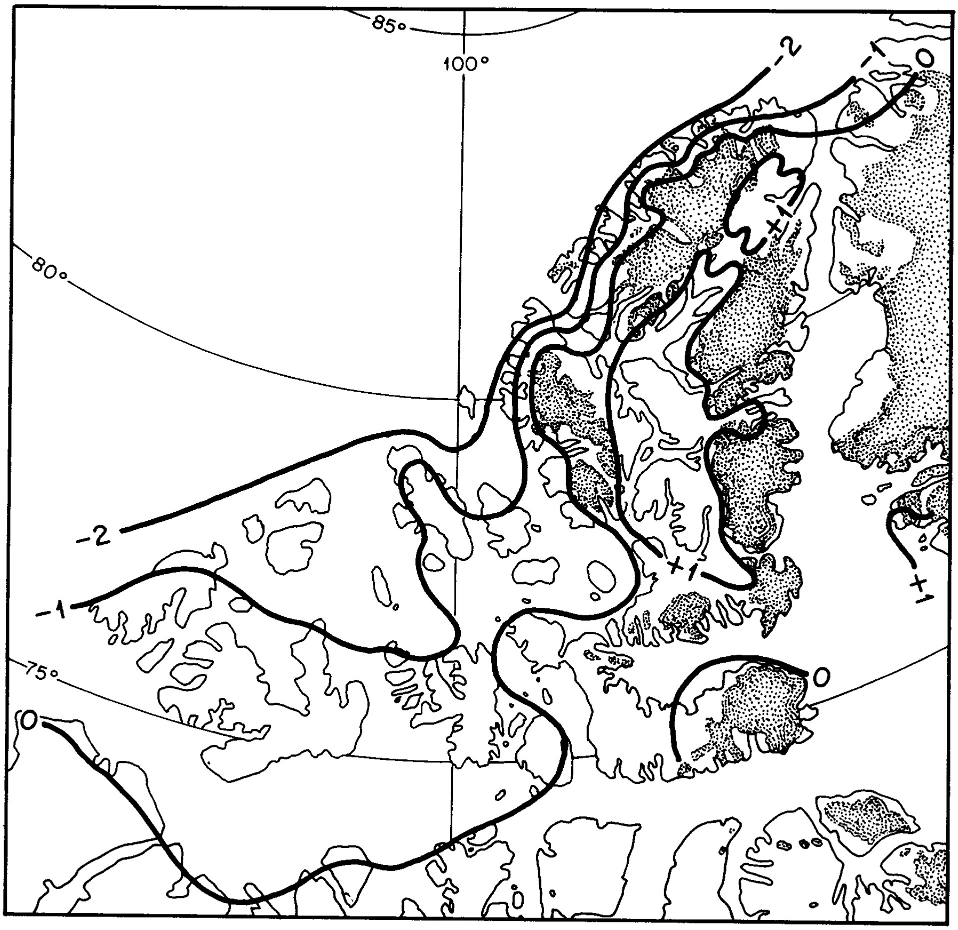

FIG. 2. Mean summer temperature, 1951–80 after Edlund and Alt, 1989).

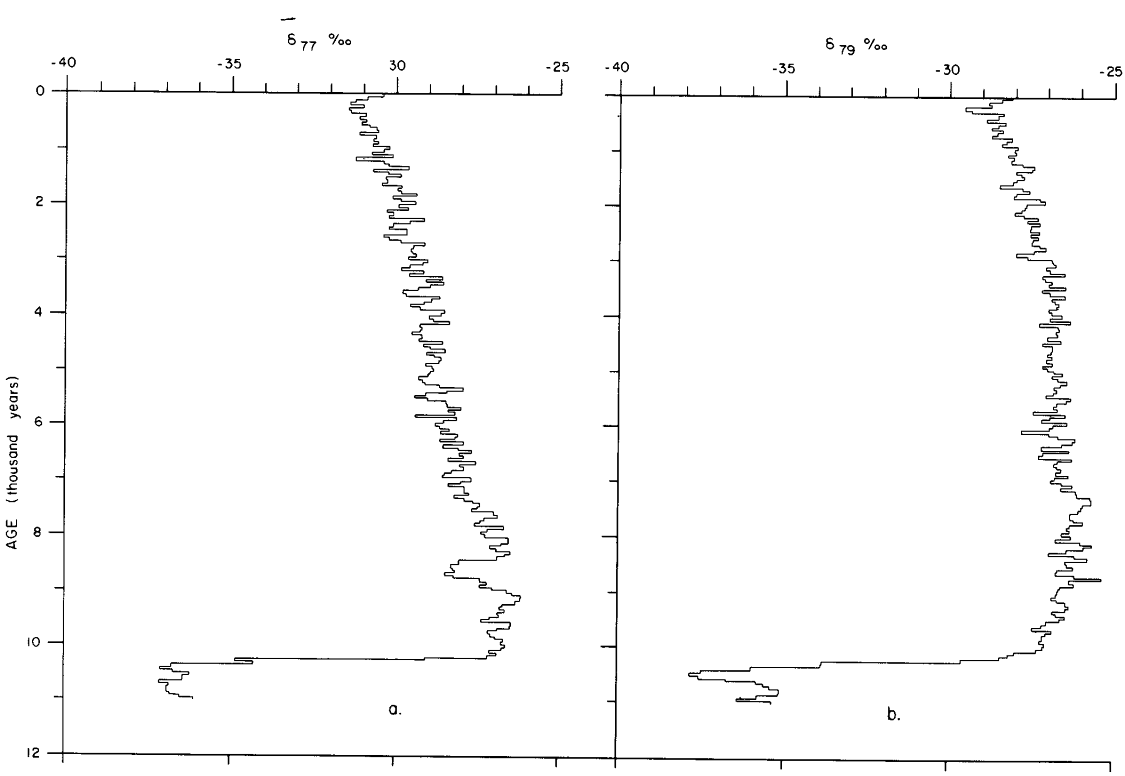

FIG 3. $\delta^{18}O$ record of the Agassiz Ice Cap ice cores (from Fisher *et al.* 1983).

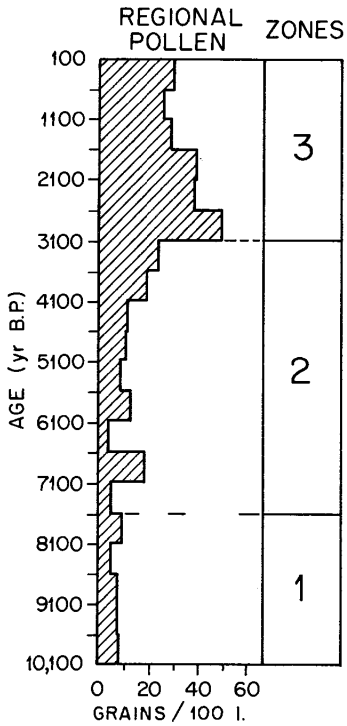

FIG 4. Regional pollen concentration in the Agassiz Ice Cap ice core A77 (after Bourgeois, 1986).

grains per 100 litres of melted ice) From 7600 to 3100 BP, total concentration increased, but largely due to more exotic pollen arriving at the site, especially *Alnus, Betula* and *Artemisia*. Regional pollen rain (i.e. from plants indigenous to the High Arctic today) did not start to increase until ~6000 BP Regional pollen continued to increase until 3100–2600 BP, declining thereafter to very low values around 250–150 BP This decline in regional pollen over the last 3000 years corresponds to the decline in $\delta^{18}O$ observed over the same interval in the core.

Of particular significance for the paleoclimatology of the High Arctic is the record of summer melt recorded in the Agassiz (A84) core (Koerner and Fisher, 1989) Summer meltwater re-freezes within the firn as superimposed ice which can be identified by the bubble content and characteristic fabric of ice at various depths in the core (cf. Koerner, 1970, 1977b) The main problem with interpretation arises because meltwater may percolate down to 'contaminate' the 'dry' accumulation of many preceding years. Nevertheless, the Agassiz core shows that remarkably high amounts of summer meltwater must have formed at the summit of the Agassiz Ice Cap during the early Holocene (Fig. 5) Temperatures must have been significantly warmer prior to 8000 BP to produce the amount of melt observed. Furthermore, apart from the present century, the last 2500 years have experienced very low amounts of melt, indicating that this has been the period of lowest summer temperatures in the entire Holocene. Interestingly, if this record is correct, it suggests that the $\delta^{18}O$ record (which shows a decline throughout the Holocene in all cores, Fig. 3) may also represent a pan-Holocene cooling trend.

Devon Island Ice Cap

The Devon Island Ice Cap (Fig. 1) has yielded three long cores from near the summit, two of which are to bedrock and may span more than 100,000 years of depositional history Although the cores are poorly dated before ~5300 (calendar years) BP, the chronology from 5300 BP to the present is considered to be accurate to within ± 10%, based on measurements of ice velocity perpendicular to the surface at various depths, seasonal cycles of microparticles (Koerner, 1977a) and on ^{14}C and ^{32}Si dates (Paterson *et al.*, 1977)

$\delta^{18}O$ values increase from pre-Holocene minima of −32 to 35‰ to mid-Holocene levels of 25.5‰ (Fig. 6A) The timing of the most pronounced change is

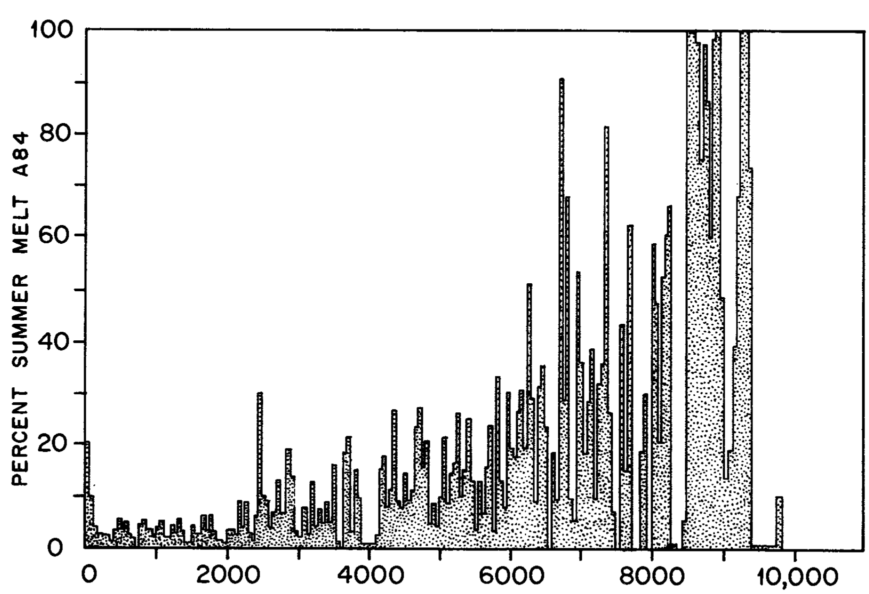

FIG 5. Percentage of melt features in Agassiz Ice Cap ice core A84 (from Koerner and Fisher, 1989).

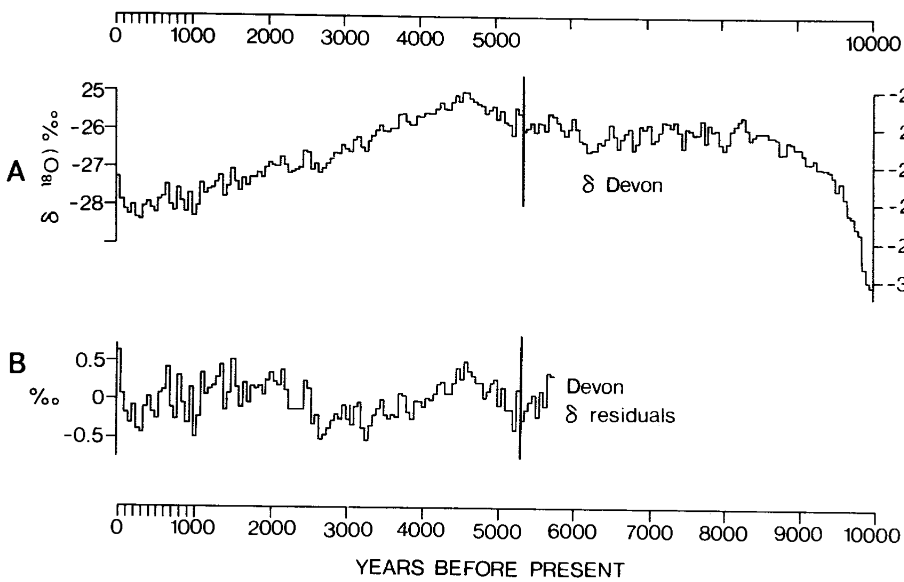

FIG 6. (A) $\delta^{18}O$ record from the Devon Island Ice Cap ice core (from Paterson *et al.* 1977). (B) $\delta^{18}O$ record from the Devon Island Ice Cap ice core after removing the low frequency trend (from Paterson *et al.* 1977).

estimated to have occurred around 10,500 BP (calendar years) based on correlations with isotopic data from Camp Century, Greenland. Part of the $\delta^{18}O$ change from pre-Holocene to early Holocene values may be due to a thinning of the ice cap by up to 150 m at the end of the glacial period. After taking into account isostatic readjustments, Paterson *et al* (1977) estimate a 7‰ 'climatic' signal at the end of the Pleistocene. $\delta^{18}O$ values then increased steadily, reaching maximum levels around 5,000 BP (calendar years) (This corresponds to ~4300 ^{14}C years BP) Since then, $\delta^{18}O$ values have decreased by 2.5 to 3‰. Part of this change may again be due to a change in the ice cap thickness, if, for example, the ice cap was thinner in the mid-Holocene and subsequently increased in size this would explain at least part of $\delta^{18}O$ change from low, early Holocene values to higher mid-Holocene values and then a further decrease to the present. Some support for this argument is provided by the fact that the $\delta^{18}O$ content of snowfall today is lower than in the mid-Holocene yet mass balance in recent years has been slightly negative If higher $\delta^{18}O$ signifies an increase in temperature (which is debatable) then there may have been a more negative mass balance in mid-Holocene times and hence a smaller (thinner) ice cap around 5000 BP Considering the probable magnitude of such changes, Fisher and Koerner (1981) estimate the decrease in mean annual temperature since 5000 BP to be ~3°C, based on the assumption that mean annual temperature decreases 1°C for a drop in $\delta^{18}O$ of 0.6‰ (Dansgaard *et al.*, 1973, Koerner and Russell, 1979) and that 2‰ of the $\delta^{18}O$ change is climatic and not due to a change in ice thickness. An alternative hypothesis is suggested by the pronounced decrease in Na and Mg content of the ice core from 3500 BP to the present. This is interpreted by Fisher and Koerner (1981) to be the result of a decrease in marine aerosol. They suggest that a reduction in storminess would reduce warm air advection (lowering the $\delta^{18}O$ content of snowfall) and also limit aerosol advection. Another explanation is that there has been more extensive sea ice since 3500 BP This would have effectively 'moved' the moisture source further from the ice cap, resulting in lower $\delta^{18}O$ levels in snow and a reduction in marine aerosols. Thus the decline in $\delta^{18}O$ may be due to a combination of lower temperatures (due to a reduction in warm air advection) *and* increased sea ice.

If the low-frequency trend in $\delta^{18}O$ is removed, the record of residuals shows two episodes of low $\delta^{18}O$ values, from ~2500–4000 BP and from ~500–100 BP, separated by periods with higher values (Fig. 6B). This pattern is also similar to that observed at Camp Century, Greenland, and shows a rough correspondence to periods of ^{14}C surplus (as recorded in tree rings) which are thought to result from reduced solar output. Fisher (1982) argues that the $\delta^{18}O$ anomalies may reflect lower temperatures resulting from reduced solar activity A reduction in the solar wind would allow cosmic rays to bombard the upper atmosphere more frequently, thereby causing more ^{14}C to be produced. In view of the complexities involved in each of these interpretations, particularly the solar activity–temperature–$\delta^{18}O$ link, the evidence can only be considered as suggestive.

The record of the last 1000 years shows very low values of $\delta^{18}O$ in the last 500 years, particularly around A.D. 1430, 1520, 1560, 1680–1730 and 1760 (Fig. 7) This compares with the lowest 'Little Ice Age' values at Camp Century around A.D 1660. (Paterson *et al.*, 1977) Higher $\delta^{18}O$ values at around A.D 1240, 1380 and 1910–1960 may indicate warmer conditions. Some support for these interpretations comes from the record of melt features — ice lenses produced by the refreezing of percolating meltwater in firn — in the Devon Island Ice Cap ice cores (Koerner, 1977b) These show, that 100–400 BP was the period with the

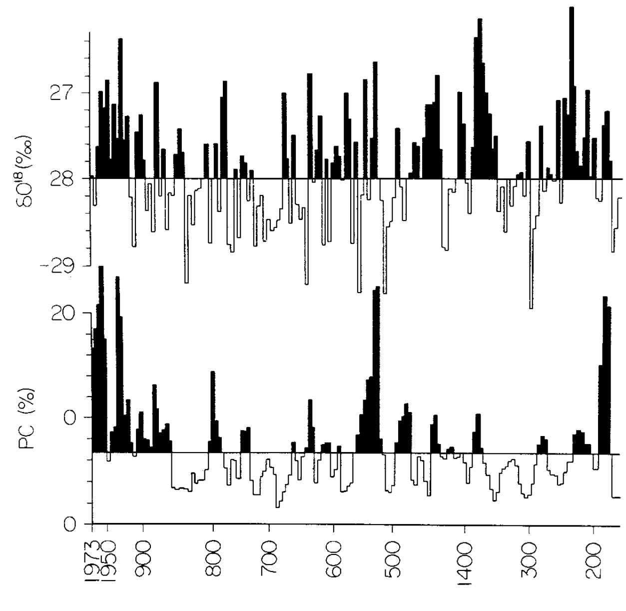

FIG 7 $\delta^{18}O$ and percentage of melt record from a Devon Island Ice Cap ice core (5 year averages for the past 500 years (Alt *et al.* 1985).

lowest percentage of melt features in the last 1000 years, following a slightly warmer period (more melt features) from 400–600 BP (Koerner and Fisher, 1985) The coldest parts of the Little Ice Age appear to have been from A.D 1680–1730 and 1820–1860 (Koerner, 1977b) Based on a comparison of both $\delta^{18}O$ and melt features, Alt (1985) argues that the period A.D 1550–1620 was a period of net *summer* accumulation at the start of the 'Little Ice Age' in the area. Because conditions on the Devon Island Ice Cap represent climatic conditions over a wide area of the Canadian Arctic Islands (Koerner, 1977b) it seems likely that this period of positive summer balance would have resulted in extensive snow accumulation over a large area of the higher mountains and plateaux of the archipelago, leading to widespread destruction of lichen and sparse upland vegetation (or simply the widespread suppression of vegetation growth, Koerner, 1980) Such areas — which can be identified today as 'lichen-free' zones — have been mapped over large areas of Baffin and Melville Islands (Andrews *et al* , 1976; Locke and Locke, 1977; Edlund, 1985)

By contrast the last 100 years have witnessed the highest levels of melt at any time in the last 1000 years (Fisher and Koerner, 1983) Summer temperatures evidently increased from 1860, reaching a maximum in the period from 1925–1960, especially in the 1930s and 1950s. This is confirmed by analysis of melt features in shorter ice cores from Ellesmere Island and Axel Heiberg Island (Hattersley-Smith, 1963, Müller, 1964, Koerner, 1977b) The increase in melt features corresponds to a period when $\delta^{18}O$ values have increased by ~1.4‰ on the Devon Island Ice Cap and 1.5‰ on the Agassiz Ice Cap, probably due to an increase in more local moisture sources (as a result of more extensive open water in the archipelago and Baffin Bay) as well as a general increase of temperature (Fisher and Koerner, 1983). Overall, the mass balance of ice caps in the Canadian Arctic Islands has been negative since at least 1925 (Koerner, 1977b, 1984)

Meighen Ice Cap

Meighen Ice Cap is a small (~85 km^2), low elevation ice cap, covering about 10% of Meighen Island. In 1965, a 121 m ice core (to bedrock) was recovered from the summit. Because the ice cap is at such a low elevation, melting takes place in most years even at the summit and superimposed ice formation is common. Consequently, seasonal variations in $\delta^{18}O$ are not visible in much of the ice core and dating by counting annual increments is not possible (Koerner *et al* , 1973) Ice fabric analysis has been used to interpret the paleoclimatic history represented in the core, sections made up of large ice crystals with few bubbles are thought to reflect warm summers (due to slow re-freezing of percolating meltwater) whereas cold summers are recognizable by smaller crystals and a high bubble content (due to rapid re-freezing of meltwater) (Koerner, 1968, Koerner and Paterson, 1974) Thus, a layer of small crystal ice from 24–44 m below the surface is considered to represent a cold period, but assigning an age to the interval is problematical. Mass balance and temperature profile measurements on the ice cap suggest that it had a negative mass balance for most of the period 1885–1965 (Paterson, 1968) Although the amount of mass lost is unknown, a maximum estimate can be made. This is based on the observation that firn is never found below 35 m in the soaked facies zone of an ice cap or glacier In the Meighen ice core, firn is found down to 22 m so there could not have been *more* than 13 m of ice lost over the most recent ablation period. Thus the fine-grained layer (at 24m) could have been as much as 37 m below the surface in 1885 Assuming a long-term balance of 12 cm (ice equivalent) per year, the time at which the cold (fine-grained) interval ended can be estimated as 308 years earlier (37/0.12) i.e. ~390 BP (Koerner, 1968) If <13 m of ice has been lost over the last century, the end of the cold period must have been later (but no later than 200 BP)

The cold interval itself spanned 20 m in the core With the assumption of 12 cm of net growth per year, Koerner and Paterson (1974) estimated that it lasted from ~390–560 BP (A.D 1400 to 1570) However, it seems likely that the net balance in this period was above the long-term mean (when ablation was higher), if we assume a net balance equal to the winter balance measured in recent years (17 cm p.a.) the cold interval may have lasted only 118 years. Clearly, changing the assumptions about the amount of mass lost this century and the net balance rate, provides considerable flexibility in estimating the time at which this cold interval began and ended. Limits seem to be about ~A.D 1400 and 1640 for the onset and 1570 to 1760 for the end.

Koerner and Paterson argue that prior to the 'Little Ice Age', from 660 BP to 2000–2500 BP, there was a long period of negative balance, now only represented by a relatively clear zone with few bubbles. The age estimates for this interval are speculative. The estimated time at which it began (2000–2500 BP) is based only on comparisons with the ice core isotopic record from Camp Century (Dansgaard *et al.*, 1971) and Nichols' (1975) reconstruction of July temperature at the northern treeline >1500 km to the south. The time at which it ended (500–660 BP) appears to be based entirely on the assumed annual balance of 12 cm p.a., with 13 m of ice lost in the last 60–80 years. Neither figure can be considered to be very reliable.

Interpretation of colder, more positive mass balance prior to this interval of mass wasting is based on more abundant dirt layers and generally low bubble content. Occasional periods of melt are suggested by layers with very large ice crystals which may represent episodes of ponding and re-freezing of meltwater on the ice surface. Dating is again speculative with an assumed time for the formation of the ice cap of 3000–4500 BP based on the evidence of lower temperatures and ice shelf formation along the northern Ellesmere Island coast (see 'Ice Shelves', below) and on the assumption that the ice cap could not have survived 'The Climatic Optimum', which (by implication) presumably occurred in the area prior to 4500 BP It is worth emphasizing that the Meighen Island ice core, by itself, does not provide evidence for such a Climatic Optimum, though many subsequent investigators have referred to Koerner's assessment of the time at which the ice cap formed as proof of a mid-Holocene 'hypsithermal' interval in the Queen Elizabeth Islands.

In summary, the dating of events represented by the Meighen ice core is mostly speculative, particularly for the period before what appears to be the 'Little Ice Age' The extent of ice loss is not known and the time at which the ice cap first developed is based only on circumstantial evidence.

Lake Sediments

Lake sediments have the potential of providing a continuous depositional record for the Holocene. However, few attempts have been made to exploit this potential. Blake has cored (mainly frozen) lake sediments at a number of sites in the area of southeastern Ellesmere Island (Blake, 1978, 1981a, 1984, 1986, 1989), pollen and diatoms from one core have been studied in detail so far (Hyvärinen 1985, Smol, 1983) Rock Basin Lake* is a small lake (150 by 250 m) in a bedrock basin above Ekblaw Glacier, which drains to Baird Inlet (Fig. 1) A near basal date of 8970 ± 160 BP at 46–51 cm depth indicates an extremely low sedimentation rate and hence points to the difficulty of a detailed paleoenvironmental reconstruction Hyvärinen (1985) distinguishes four pollen zones which are primarily successional in nature, not climatic. However, he does note that the uppermost pollen zone (zone 4: ~3500 BP to present) may be climatically significant because around this time a period of relative vegetational stability came to an end. Plant communities more characteristic of open soils and fell fields increased in importance Drier and poorer heath-type vegetation may have spread, with an increase in the presence of mosses in heath communities. Hyvärinen notes that such data only provide a minimum estimate on any associated climatic change because the site may be insensitive (complacent) to a shift in climate, and a deterioration may have begun before it was eventually manifested in the pollen record. In this regard it is worth noting that the general decline in pollen concentration actually began between the levels dated 6780 ± 220 and 4080 ± 210 BP

In a study of diatoms and chrysophytes from the same core, Smol (1983) concludes that Hyvärinen's zones 2 and 3 (~7500? to 4000? BP) represent the period of maximum warmth (Fig. 8) This conclusion is based on the argument that warm summers lead to a reduction in the amount of lake ice, enabling relatively deep-water diatoms to proliferate (Smol, 1988) In cold periods, only a marginal lead is exposed so that only shallow water diatoms can survive In Rock Basin Lake, a shift to a more diverse assemblage of shallow water diatoms and an overall decline in total abundance after ~4000 BP suggests that summer temperatures declined at about this time, and no subsequent recovery is apparent in the diatom record. Some support for warmer conditions in the mid-Holocene is provided by Blake (1989) who reports on radiocarbon-dated algae (*Mougeotia* spp.) from sediments at the bottom of a small lake, situated on a hill summit at the edge of the ice cap southeast of Bay Fiord (Fig. 1d). The lake ice is currently over 5 m thick in places and is completely frozen to the underlying sediments. The algae dated 5730 ± 70 BP, indicating warmer conditions at that time, with thinner ice cover enabling the algae to survive However, when such conditions began, or how long they lasted, cannot be determined from this one sample

Lake sediments from northeastern Ellesmere Island have been studied by Retelle (1986), although these provide a record of post-glacial emergence along the coast (Retelle *et al.*, 1989), sedimentological changes provide only a weak paleoclimatic signal. They suggest the presence of landfast sea ice or ice shelves in the early Holocene (cf. England, 1983), more open sea ice conditions and increased runoff in the mid-Holocene and cooler conditions with reduced runoff in the late Holocene Pollen analysis revealed no meaningful paleoclimatic signal (Blackman, unpublished manuscript)

One other aspect of lake sediment studies worth noting is the fact that few sediments have been recovered which date before ~9000 BP Blake (1981a, 1987a) reports basal dates of 9370 ± 110, 9130 ± 160 and 8970 ± 190 BP in cores from lakes on southeastern Ellesmere Island. Blake notes that these dates are similar to the oldest dates on peat deposits from the Carey Islands, which dated 8940 ± 90 BP (Brassard and

*Place names mentioned may not be officially recognized by the Canadian Commission on Geographic Names. However the names used here are those given in the papers referenced.

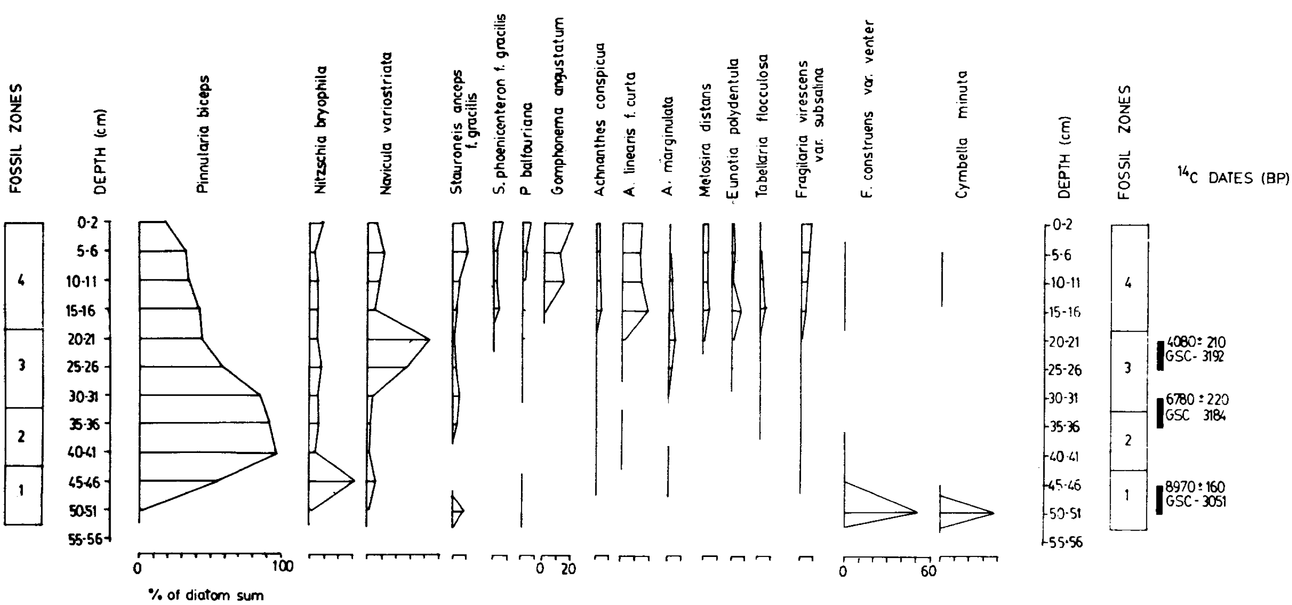

FIG 8. Diatom record from Rock Basin Lake (from Smol, 1983).

Blake, 1978) and there are few older dates, even considering all the lakes studied on Greenland where the oldest dates reported on lake sediments are 10,050 ± 150 BP in Scoresby Sund, east Greenland and 9840 ± 170 BP in southwestern Greenland (Kelly and Funder, 1974) However, the significance of this 'absence' of older sediments is not clear; sub-bottom profiles have not been reported from any of these lakes, to demonstrate that the total sediment thickness in each of the basins has been sampled. Furthermore, most of the lakes studied by Blake are within the most glacierized region of Ellesmere Island today so that even if the lake basins in this area became ice-free and began accumulating sediment around 9000 BP, this may not be of great significance for other parts of the archipelago Some areas of Ellesmere Island may have been ice-free for 30,000 years or more (England and Bradley, 1978) so that the recovery of long lacustrine sedimentary records from these areas may be possible. Indeed, pre-Holocene radiocarbon dates have been reported for sediments from Beaufort Lakes, northeastern Ellesmere Island, though they have been interpreted as probably contaminated with 'old carbon' and hence erroneous (Retelle, 1986a, Retelle *et al* , 1989).

Peat Deposits

Although peat can be found throughout the High Arctic, accumulations are rarely very thick and generally occur only in isolated patches. Ovenden (1988) argues that the timing of the onset and cessation of peat growth is of climatic significance, her compilation of dated peat sections reveals maximum peat growth episodes in the mid-Holocene on Ellesmere Island and early to mid-Holocene on islands to the south and west. However, these conclusions are based on a questionable assumption, for sections which are not bounded by uppermost and lowermost dates, Ovenden arbitrarily assumes peat growth ceased 2000 years after the onset of peat accumulation (if this is dated) or began 2000 years prior to the end of peat accumulation (if this is dated) Further analysis of the dated sections indicates a mean peat accumulation rate for High Arctic sites of 1 mm per year in sections where two or more dates enable such rates to be calculated. Applying this rate to all other sections provides the age/frequency histogram shown in Fig. 9 A possible interpretation of these data is that peat accumulation was increasingly common in the High Arctic landscape from the early Holocene to ~3500 BP, but its occurrence subsequently declined. However against this must be weighed the natural tendency to collect and date only thick peat sections, which are uncommon. The average thickness of dated sections from the High Arctic compiled by Ovenden (1988) is 205 cm and the average duration represented is 2580 years (assuming 1 mm per year accumulation when not otherwise known). Consequently, there is a sampling bias against 'young' peat sections (<200 cm, <2500 years) which makes it difficult to determine if there has been a real decline in peat accumulation in the late Holocene, or not. In fact, the whole notion of interpreting the onset and termination dates of peat growth in terms of climatic conditions either 'favorable' or 'unfavorable' for peat growth is of dubious validity

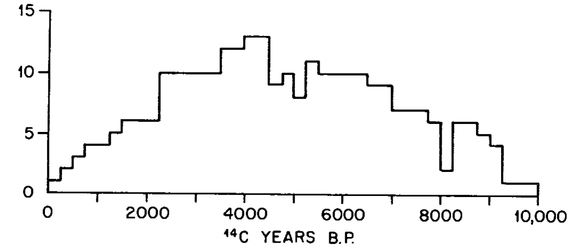

FIG 9. Frequency count of the number of occurrences of dated peat, accumulating during each 250 year interval, assuming 1 mm per year growth when not otherwise known (data from Ovenden, 1988).

Peat accumulation is very much dependent on topography, specifically locations where waterlogged conditions can prevail year after year Arctic landscapes are notoriously unstable and normal surface instability tends to create only relatively short-lived environments favorable for peat growth. For example, many dated peat sections are from high-centered polygons where natural evolution of the system tends to lift and expose peat which formerly accumulated in peripheral, waterlogged environments. The fact that the average period of time represented by dated High Arctic peat sections is 2600 years clearly indicates this instability Peat should perhaps be considered as accumulating in a mosaic of transient environments, clearly controlled by climate at the large scale, but present in specific locations as a result of local topographic and hydrological factors, the evolution of which may be unrelated to climate (cf. Jacobs and Leung, 1981)

Few stratigraphic palynological studies of peat deposits have been carried out; those that have show little change in pollen spectra through time (e.g. Bliss, 1975, Jankovska and Bliss, 1977) A peat section in front of the Thompson Glacier, Axel Heiberg Island revealed very minor peaks in Cyperaceae, Graminae, *Salix* and *Saxifrage* pollen at a level estimated to be between 2000 BP and the base of the section (dated at 4210 ± 100 BP) (Hegg, 1963) Hegg argues that these slightly higher pollen counts indicated a climate 'more favorable than at present', but the evidence is not very compelling; indeed, Hegg notes that pollen content may reflect changes in water level, perhaps related to topographic changes and not climate. Nevertheless, Hegg's conclusion is in agreement with pollen evidence from the Agassiz Ice Core (discussed earlier) in which maximum regional pollen concentrations occurred between 3100 and 2600 BP (Bourgeois, 1986)

Driftwood

Driftwood can be found at various elevations above present sea level throughout the Canadian Arctic Islands. The wood originates in forested areas of northern Canada, Alaska and the Soviet Union, from there it is carried to the Arctic Ocean where it passes through ice-choked seas and channels, eventually to be deposited on the shores of the islands. Because all islands of the Canadian Arctic archipelago have experienced glacio-isostatic emergence, driftwood deposited in this way can be found at all elevations up to the marine limit. Age of the wood can be obtained by ^{14}C measurements on the wood itself, or by estimation (based on the elevation of the wood) if local emergence rates are known (Stewart and England, 1983) The latter approach may often underestimate age, as wood samples are commonly displaced downslope (to 'younger' surfaces). Data on the frequency of occurrence of wood of various ages (mainly from southern Ellesmere Island) led Blake (1972) to argue that the greater abundance of driftwood between ~4500–6500 years BP reflected more open water conditions at that time, which enabled driftwood to reach the coast more easily A larger sample of driftwood dates from the High Arctic and northern Greenland led Stewart and England (1983) to a similar conclusion (maximum driftwood penetration ~4200–6000 BP if probable downslope movement of samples is taken into account) There can be little argument that the driftwood collected so far reveals a maximum age–frequency distribution in the 4000–6000 BP range (Fig 10) What is less clear is the paleoclimatic significance of this fact. Blake (1972) notes that the amount of driftwood on a shore is a function of several variables. distance from source of supply, duration and extent of open water (the paleoclimatic signal), the nature and exposure of the coastline, and the rate at which land is emerging or submerging. To these should be added the rate at which driftwood is being produced and/or carried northward to the Arctic Ocean Although trees were present in the Mackenzie drainage in the early Holocene (Ritchie and Hare, 1971) for boreal forests as a whole, the mid-Holocene was a period of maximum expansion to the north (e.g. Khotinsky, 1984, Ritchie, 1984) It might therefore be argued that the input of wood to the Arctic Basin would have reached a maximum at that time, thereby increasing driftwood flux into the Canadian Arctic archipelago and resulting in a depositional maximum, regardless of any change in *local* ice conditions. The significance of driftwood is further confounded by Häggblom's observations of maximum driftwood occurrence on Spitsbergen which are interpreted as reflecting *less* open water and *more* sea-ice (Häggblom 1982) It seems impossible to reconcile Blake's and Stewart and England's interpretation with that of Häggblom and only additional supporting data can provide confidence in the interpretation of driftwood data from each area. Stewart and England (1983) for example, note that shells of the

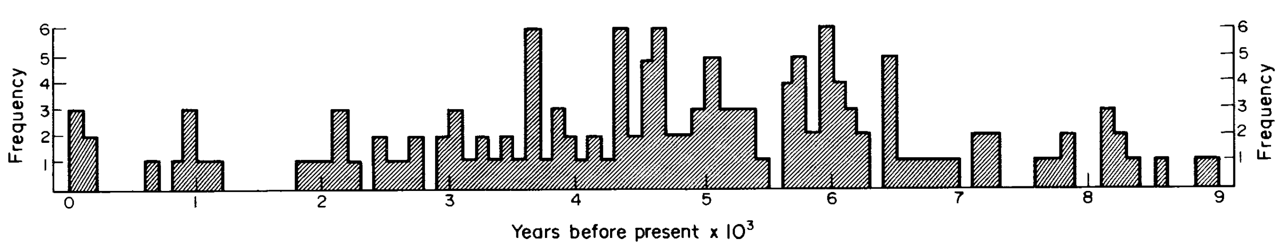

FIG 10. ^{14}C age distribution of dated driftwood samples from the Queen Elizabeth Islands and northern Greenland (basic data from Stewart and England (1983), with additional samples listed in Lemmen 1988) Blake (1987b, 1988), and Evans (1988).

subarctic-boreal mollusc *Limatula subauriculata*, currently only found south of 73°N, were recovered from raised marine deposits in Clements Markham Inlet, northern Ellesmere Island (~82°N) These shells dated 6400 ± 60 BP which indicates that warmer water conditions (and by inference, less sea ice) prevailed at very high latitudes around the time that driftwood abundance began to increase. Additional support for extensive open water at this time comes from a narwhal tusk found on the northwestern coast of Ellesmere Island, behind a present-day ice shelf, which dated 6830 ± 50 BP (Evans, 1989) This is discussed further in the next section

Ice Shelves

The only place in North America where ice shelves currently exist is along the northern and northwestern shores of Ellesmere Island (Fig. 1) Ice shelves form by the accumulation of snow and superimposed ice on top of coastal fast ice (sea ice) together with re-freezing of freshwater on the bottom of the ice. At the present time, the Ward Hunt Ice Shelf — the largest on northern Ellesmere Island — is over 50 m thick in places (Jeffries, 1987) Periodically, large sections may become separated and drift into the Arctic Ocean where they are designated as ice islands. Of the many ice islands which have been occupied by scientific parties, T-3 has been studied the most (Crary, 1956, 1960) Cores taken from T-3 revealed an uppermost dirt-laden layer overlying ~30 m of clean ice, then a very dirty layer with 10 m or more of iced firn and sea ice below. Radiocarbon dates have been obtained on bulk samples from the upper and lower dirt layers, though their meaning is not clear If they are indeed composed of wind-blown material from the adjacent coast, presumably the material could contain older organics which would contaminate the sample and so provide only a maximum age for the layer This seems to be confirmed by dates of 5720 ± 250 and 5830 ± 200 years BP on the uppermost dirt layer, dates of 4370 ± 200 and 4480 ± 200 years BP *below* this, and 3050 ± 150, 3100 ± 180 and 3000 ± 200 years BP on the lowermost dirt layer (Crary, 1960) In spite of these inconsistencies, Crary argues that the date of 3050 ± 150 BP is the most reliable estimate on the age of the lower dirt layer Largely on this basis, Crary *et al.* (1955) suggest that the ice shelf, from which T-3 derived, developed in the cool period after the 'Climatic Optimum' (the age, and even the existence of which is not known in this region, Crary *et al* assume that this 'Optimum' occurred 5000–6000 BP, with ice shelf formation 3500–5000 BP) They further claim that the lower dirt layer represents a prolonged period of ablation in Medieval times followed by a period of accumulation (the upper clean ice layer) and then ablation in the last 200 years, leading to the surface dirt layer All of these interpretations must be viewed as speculative, based largely on the assumption that climatic changes observed in western Europe — the Hypsithermal, Medieval warm epoch etc. are relevant to northern Ellesmere Island. The ice shelf stratigraphy and associated radiocarbon dates provide no confidence that such a scheme can be applied to this region, and further studies of the problem from new ice shelf cores are clearly warranted.

Perhaps of more direct relevance to climatic conditions along the shores of the Arctic Ocean are radiocarbon dates on driftwood found behind present-day ice shelves (Fig. 11). These must have been washed ashore during periods when the ice shelves were not blocking the northern beaches and fiords, as they are today The dates indicate that from ~6900 BP (or earlier) to ~3000 BP (or later) driftwood was able to penetrate to locations behind Ward Hunt Ice Shelf (Crary, 1960; Jeffries, 1987, Lemmen, 1988) Furthermore, marine shells on a raised beach 38 m above sea level on Ward Hunt Island dated 7200 ± 200 BP (re-dated by Lyons and Mielke (1973) at 7755 ± 150 BP) suggest that the ice shelf may also have been absent well before 6600 BP Indeed, driftwood found 3 km east of the Cape Discovery Ice Rise (behind Ward Hunt Ice Shelf) was dated at 8850 ± 50 BP (Lemmen, 1988). There are also reports of driftwood from Yelverton Inlet (behind the ice shelf in the western fiord arm, Fig. 1b) dating 6410 ± 250 and 8150 ± 140 years BP (Blake, 1972) and 8120 ± 80 (Blake, 1987b), and Evans (1989) reports the unique find of a narwhal tusk, dating 6830 ± 50 years BP behind the present day Cape Alfred Ernest Ice Shelf (Fig. 1b) Narwhals have rarely been sighted beyond areas of relatively open water, indicating that extensive ice-free conditions existed in the Canadian Arctic archipelago at this time Together with driftwood from the same area dating 4310 ± 70 and 7450 ± 80 years BP (Evans, 1988) these samples support the argument that ice shelves were largely absent from the northern and northwestern coasts of Ellesmere Island for most of the Holocene

Additional dates on macrofossils collected from the

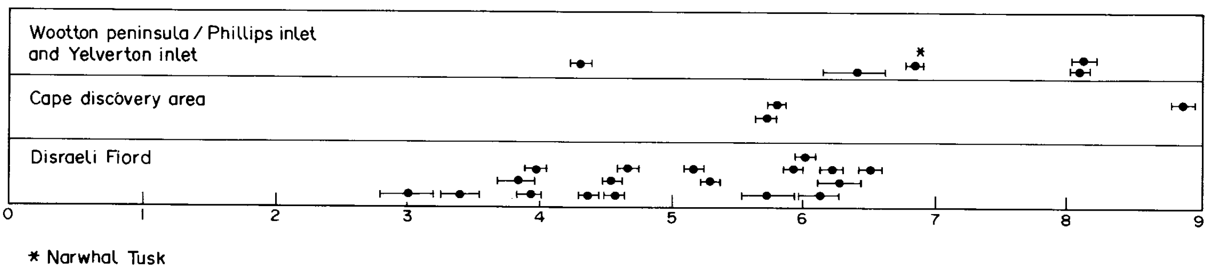

FIG 11 ^{14}C dates on driftwood from behind ice shelves on northern Ellesmere Island (see text for data sources).

surface of Ward Hunt Ice Shelf provide further insight into the timing of ice shelf formation. Shells and sponges, considered to have been frozen into the base of the ice shelf and carried upwards to the surface by ablation at the top and further freezing on the bottom, have provided ^{14}C dates of 13200 ± 440, 6815 ± 190, 3680 ± 100, 3645 ± 120, 3400 ± 140 and 3390 ± 130 BP Although these dates, individually, are of little paleoclimatic significance, it is of interest that none of the samples dated <3400 ± 140 BP, suggesting that the ice shelf itself was not present before that time (Lyons and Mielke, 1973) This approximates the age of the youngest driftwood found behind Ward Hunt Island (3000 ± 200 BP) which suggests that ice shelf growth began sometime after 3000 BP However, Lyons and Mielke (1973) suggest ice shelf growth began earlier, around 4100 BP based on a supposed cool period recorded in the Camp Century ice core and Hattersley-Smith's (1969) evidence of 'glacial readvances within the last 4000 years' They argue that driftwood in the 3000 to 4000 year BP range may have reached the shore via a lead behind the incipient ice shelf, but there is no strong evidence in favor of this and it seems more logical to accept that ice shelf formation occurred after 3000 BP

It is worth noting that ice shelves may develop fairly rapidly; Crary estimates that even with 10 cm of annual ablation, a 50 m ice shelf can develop in 360 years, and up to 11 m of new ice growth in 14 years has been observed in historic time (Lyons and Ragle, 1962) Furthermore, ice shelf break-up in historic time has not been clearly related to climatic conditions and so absence of ice shelves may relate to some mechanical disruption (tidal, seismic) which is of no climatic significance (Holdsworth, 1971, 1977, Jeffries, 1987) Hence, ice shelves may be transient phenomena which have developed and broken up on numerous occasions in the past and they may not have been present simultaneously at all locations where they are found today

In summary, no driftwood dating between ~8000 BP and ~7000 years BP has yet been found behind modern ice shelves, so ice shelves may have been present at this time, broken up between 7000 and 2800 years BP, and subsequently re-formed. The problem is that evidence for ice shelf *presence* is based largely on evidence of driftwood *absence*, and there have been few systematic searches along coastal areas behind the ice shelves to collect and date driftwood (although many samples which have been collected have yet to be dated, Stewart and England, 1983, Lemmen, 1988) As new driftwood dates are reported, it is likely that our understanding of ice shelf dynamics and their climatic significance will change

Glacial Geologic and Glacio-Isostatic Evidence

Indirect evidence of paleoclimatic conditions is provided by glacial geologic and glacio-isostatic evidence pertaining to the timing of deglaciation in the region, though regional differences in the timing of deglaciation (in the late Pleistocene or early Holocene) may occur due to factors of ice dynamics rather than climate. Eustatic effects on calving glaciers (resulting from ice volume changes, far from the High Arctic) may also play a role in the regional pattern of deglaciation. Furthermore, geomorphic features relating to initial deglaciation may not be formed or preserved, and may not contain dateable material so it is necessary to adopt a large-scale, regional approach to understand paleoclimatic events from this evidence.

Figure 12 shows the pattern of deglaciation of the region, from 10,000 BP to 7000 BP, according to the composite reconstruction of Dyke and Prest (1987) Many islands in the western and southwestern part of the archipelago do not appear to have been covered by ice caps in the late Pleistocene, if they were, the ice must have been thin and stagnant, leaving essentially no geomorphological evidence

Deglaciation was most rapid between 9000 and 10,000 BP in the south and west with the disappearance of the Bathurst Island Ice Cap by 9000 BP Farther north and east, deglaciation began around the beginning of the Holocene in many areas, possibly accelerated by eustatic sea-level rise Minimum estimates on the timing of initial post-glacial emergence are provided by samples dating 10,100 ± 210 BP near Alert (England, 1978), 9560 ± 70 BP in the Marvin Peninsula area (Lemmen, 1988), 9845 ± 485 BP in Clements Markham Inlet (Bednarski, 1986) and 10,140 ± 90 BP at Cape Armstrong (Philips Inlet area) (Evans, 1988) On northeastern Ellesmere Island (Kennedy and Robeson Channel area) dates of 8100 to 8600 BP also provide minimum estimates on initial emergence (England, 1982 1983, Retelle 1986) Further south, in Otto Fiord, the earliest dates on deglaciation are ~9100 BP (Bednarski, 1987) and in west-central Ellesmere Island, around 8000 BP (Hodgson, 1985, England, 1987) On Devon Island deglaciation was in progress by 9500 BP (Barr, 1971)

In some areas, uplift following emergence was initially quite slow; Lemmen (1988) for example demonstrates that rapid retreat of glaciers in the Marvin Peninsula area (northern Ellesmere Island) did not really get underway until ~8100 BP which is similar to events further east in Clements Markham Inlet, documented by Bednarski (1986, 1988) Farther south, glaciers retreated rapidly in the early Holocene and, in fact at least one glacier in southeastern Ellesmere Island was behind its present day terminal position by 7700 BP (Blake, 1986, see discussion below) By contrast, England (1978) and England and Bednarski (1986) report that glaciers on the Hazen plateau (northeastern Ellesmere Island) were still within a few kilometers of their maximum (late Wisconsinan) position until ~6200 BP This seems likely to be related to dynamical (lag) effects rather than reflecting a distinctly different climatic record for this area.

Overall, glacial geologic and glacio-isostatic evidence indicates a rapid amelioration of climate in the early

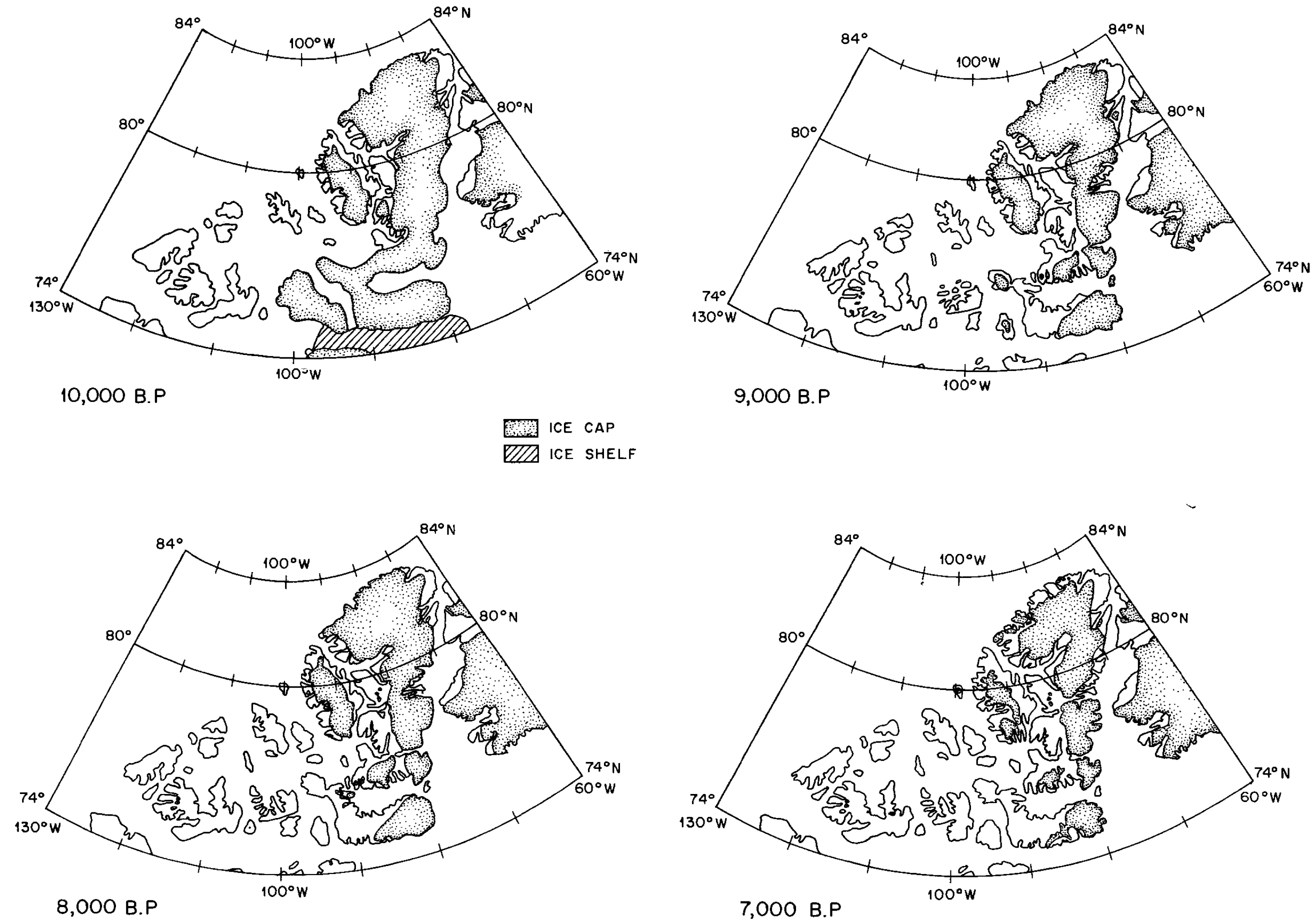

FIG 12. Ice cover over the Queen Elizabeth Islands at 1000 year intervals from 10,000 to 7000 years BP (after Dyke and Prest, 1987).

Holocene, with the southwestern part of the archipelago losing its limited ice cover fairly quickly and the higher elevation regions to the northeast and east responding more slowly This probably reflects the inertial effects of the mountain-based ice sheets in these areas, down-wasting in response to the same climatic amelioration but not undergoing significant terminal recession until 1000–2000 years later

Glacier Margins

In many locations throughout the glacierized regions of the High Arctic, samples of organic material (generally peat, driftwood or marine pelecypods) overlain by, or incorporated into till or glacial outwash attest to the former extent of ice. The most dramatic example is from Axel Heiberg Island where driftwood from beneath the Thompson glacier, 1.5 km from the snout, dated 5480 ± 100 and 5920 ± 100 years BP, clearly indicating a significantly smaller glacier at that time (Müller, 1964). Driftwood from innermost Disraeli Fiord (northern Ellesmere Island) 11 km inside the floating glacier margin of the present Disraeli Glacier, dated 6030 ± 70 BP indicating far less ice in the area in the mid-Holocene (Lemmen, 1988) Similarly, Blake (1985) reported the age of pelecypod fragments and barnacles found along the northern side of Macmillan Glacier, Baird Inlet (Ellesmere Island) to be 7670 ± 100 and 7240 ± 60 years BP, also indicating reduced ice extent in this area. In the same region shells and calcareous algae incorporated into a lateral moraine on the south side of Leffert Glacier date 2280 ± 140 to 2880 ± 70 BP, indicating ice has advanced at least 7 km within the last 2000 years (Blake, 1984, 1986)

Such findings, providing clear evidence of smaller glaciers in the mid to early Holocene, are rare. More commonly, organic material immediately in front of present day glacier termini have been dated to provide evidence that modern glacier positions have not been exceeded for a particular length of time. For example, peat ~2 km from the terminus of the Thompson Glacier, Axel Heiberg Island, dated 4210 ± 100 BP, indicating the ice front has advanced no more than 2 km beyond its present position in the last 4200 years. Even more limiting is the date of 8930 ± 100 BP on marine pelecypods within 15 m of the present margin of Alfred Newton Glacier (southeastern Ellesmere Island) indicating the ice is as far advanced as at any time in the last 9000 years (Blake, 1981b) Other evidence includes:

(1) on northeastern Ellesmere Island, the Jens Glacier is advancing over raised marine deposits dated 6595 ± 280 BP (England, 1978),

(2) at the northern end of Makinson Inlet, south-

eastern Ellesmere Island, a thick (8–9 m) peat section in front of glacier 7A-45 dated 5180 ± 260 to 2590 ± 150 BP (Blake, 1981b);

(3) marine shells within 500 m of the present ice margin near the head of McClintock Inlet, northern Ellesmere Island, dated 7770 ± 70 BP (Lemmen, 1988),

(4) peat 100 m in front of the Webber Glacier (Borup Fiord area, northwestern Ellesmere Island) dated 4190 ± 130 BP (Dyck and Fyles, 1963),

(5) organic material within 400 m of the present terminus of Air Force Glacier (near Tanquary Fiord, Ellesmere Island) dated 930 ± 140 BP, and plant material from near the snout of the Gilman Glacier, northern Ellesmere Island, dated 965 ± 75 BP (Hattersley-Smith, 1969, 1972)

All these dates point to two conclusions. firstly, glaciers throughout the region are currently more extensive than for at least several thousand years, and in many cases are as far advanced as at any time since the end of the last (Wisconsinan, or Foxe) glaciation (cf. Miller, 1976) Secondly, limited evidence points to a significant *recession* of ice behind current positions from 5400 to 6000 years BP (or earlier) on Axel Heiberg Island, around 6000 BP on northern Ellesmere Island and 7750 to 7200 years BP (or earlier) perhaps continuing to 2200 BP on southeastern Ellesmere Island. At what point in time glaciers re-advanced is not known with certainty, but several lines of evidence provide some estimates. On Ellesmere Island, advancing glaciers cut off a part of a fiord to create Lake Tuborg, now a meromictic lake containing 'trapped' seawater Because the seawater was isolated from interaction with the atmosphere, a radiocarbon date of 3000 BP on bicarbonate precipitated from samples of the water is thought to provide a minimum date on the timing of the ice advance (Long, 1967) Similarly, 1 km in front of glacier 7A-45 (southeastern Ellesmere Island, Fig. 1d), peat which dated 2590 ± 150 BP (10 cm below the top of the section) is overlain by up to 1 m of coarse outwash and bouldery till, indicating an ice advance after 2500 BP and subsequent recession. At nearby Hook Glacier, willow twigs incorporated in till near the glacier terminus dated 1200 ± 200 and 1110 ± 60 BP, indicating ice advanced there some time after 1100 BP (Blake, 1981b) Near Borup Fiord, 400–500 m in front of the Carl Troll Glacier, till overlies organic material dated at 2900 BP, indicating ice advanced after that date and has subsequently retreated (Völk, 1980; King, 1981) On northwestern Ellesmere Island, Alfreds Glacier (Philips Inlet area, Fig. 1b) has advanced since 1850 ± 50 BP, thrusting upwards blocks of glacimarine silts containing valves of *Astarte borealis* (Evans, 1988) Finally, peat beneath 30 cm of 'glacial outwash and morainic material' near the eastern edge of the Eugene Glacier, northern Ellesmere Island, was dated at 4900 ± 60 BP, indicating a reduction in the size of this piedmont glacier in the mid-Holocene and a subsequent re-advance (Lowdon and Blake, 1979)

These observations suggest that glaciers re-advanced between 1000 and 3000 years BP following a mid- to early Holocene period of reduced ice extent. In many areas this advance brought glacier termini to the most advanced positions they have occupied in the entire Holocene.

The final set of evidence derived from glacier positions comes from ^{14}C dates on organics exposed by recent ice recession. In many locations, plants (often in seemingly undisturbed growth position) are exposed as present-day glacier termini undergo recession. Bergsma *et al* (1984) for example, describe well-preserved *Dryas integrifolia* cushions, with flower heads and seeds attached, and *Salix arctica*, with foliage attached, being revealed at the snout of the receding Twin Glacier, Alexandra Fiord, eastern Ellesmere Island. ^{14}C dating gave ages of 400 ± 140, 430 ± 90 and 410 ± 45 years BP, indicating the plants were buried by snow and ice between A.D. 1410 and 1690 (taking the ^{14}C dates at face value, not accounting for variations in ^{14}C production rates, cf. Stuiver, 1982) Similarly, Müller (1964) reports that plants dated at 240 ± 100 BP, ~50 m in front of the White Glacier, Axel Heiberg Island, had been over-ridden in the most recent ice advance Comparable evidence is described by Blake (1981b) who reports two dates from central and southern Ellesmere on recently over-ridden organic material. 150 ± 50 BP (Sverdrup Pass) and 120 ± 50 BP (Sydkap Ice Cap) All these lines of evidence clearly indicate that glaciers advanced within the last few hundred years (120 to 450 BP) Although glaciers are generally receding in most areas today (cf. Hattersley-Smith, 1972, Bergsma *et al* , 1984, King, 1983, Evans, 1988) in many cases ice positions are still as far advanced as at any time since the last glacial maximum.

Expansion of the Ranges of Plants and Animals

A variety of fossils provide evidence that many species of plants and animals were formerly more widely distributed in the archipelago than they are today In most cases, they indicate warmer temperatures, and less severe environmental conditions in the early to mid-Holocene One such line of evidence has already been discussed — a narwhal tusk dating 6830 ± 50 BP found behind the modern Cape Alfred Ernest Ice Shelf on northwestern Ellesmere Island Fossils of other marine mammals which generally only frequent areas of relatively open water, have also been collected. Harington (1975) reported on walrus bones found on Bathurst Island dating 7320 ± 120 and a bowhead whale bone dating 7380 ± 140 found on Cornwallis Island. A walrus tusk from Coburg Island (at the eastern end of Jones Sound, Fig. 1d) dated 8690 ± 100 and whale ribs found on eastern Ellesmere Island (Baird Inlet area) dated 6500 ± 260 and 6920 ± 90 BP (Blake, 1987b). Recently, Bednarski (1989) has described bowhead whalebones from the Nansen Sound coast of Axel Heiberg Island, dating 7475 ± 220 BP Relatively warmer waters are also suggested by the occurrence of the sub-Arctic pelecypod *Limatula sub-*

auriculata in the northern fiords of Ellesmere Island at 6400 ± 60 BP (Stewart and England, 1983) Together with the evidence of driftwood occurrence along the northernmost shores of the region as early as 8915 ± 115 BP (Stewart and England, 1983) it seems that summer sea-ice conditions in the early Holocene were at least as open, and probably much more open, than is common today

On land, a few lines of evidence also point to milder summers in the early Holocene. On Skraeling Island (Alexandra Fiord, east-central Ellesmere Island) peat dating 6650 ± 70 BP contained *Potamogeton filiformis*, a plant which is currently only found in the low Arctic (over 1000 km farther south) where summer temperatures today are ~2–3°C warmer (Blake, 1982, Edlund, 1986). Fragments of the same plant, together with *Betula* seeds and low Arctic insect fragments were also found near Canon Fiord, west-central Ellesmere Island and may represent conditions around 8000 BP (Hodgson, 1985). However, there is a possibility that the deposit represents an earlier interglacial or interstadial period (or is contaminated with older material) since *Salix* twigs in the sample dated >16,000 BP and >38,000 BP

On southern Bathurst Island, peat containing *Vaccinium uliginosum* and *Salix polaris* ssp *pseudopolaris* dated 9210 ± 170 BP (Ovenden, 1988), these woody plants are not found today in the region except along the extreme southern peninsula of Melville Island (Edlund, 1983) Their occurrence only shortly after deglaciation of the area indicates rapidly ameliorating conditions very early in the Holocene, to levels which are even more favorable than today Finally on Ellef and Amund Ringnes Islands, D Hodgson obtained peat dating 7500 ± 90 and 6900 ± 100 BP, containing *Salix* and *Carex*, both of which are virtually absent from the area today (reported in Ovenden, 1988) One other report of related interest is evidence that caribou were present on the northern coast of Ellesmere in the early Holocene (Stewart and England, 1986) A caribou antler from Clements Markham Inlet dated 8415 ± 135 BP, indicating that conditions on land (at 82.5°N) were productive enough in the early Holocene to support large herbivores A similar find on Peary Land (northern Greenland) dated 8750 BP; together, these finds strongly suggest a productive ecosystem at the highest latitudes by the early Holocene Indeed, it may be that coastal areas which were not inundated by ice in the last glaciation acted as biological refugia for some plant and animal species, enabling rapid expansion of populations as climatic conditions improved and more areas became free of snow and ice (cf. Leech, 1966; Brassard, 1971, England and Bradley, 1978) However, as yet there are no dated animal fossils to support this notion.

DISCUSSION

From the wide variety of evidence discussed above, four broad conclusions can be derived. These are discussed in turn, below·

Summer temperatures in the early to mid-Holocene were probably as high as they have been for much of this century, perhaps even higher

There is considerable evidence for a mid-Holocene warm period, from ~4000–6000 ^{14}C years BP This includes. maximum driftwood occurrence throughout the archipelago, driftwood found behind areas currently isolated from the sea by ice shelves, maximum regional pollen concentrations in ice core and lake sediment records, and maximum δ^{18}O values in the Devon Island Ice Cap ice core (at ~4300 ^{14}C BP). There is also evidence, however, that the warmest period may have been even earlier In the Agassiz ice core, maximum melting took place *before* 6000 ^{14}C years BP (~6700 calendar years) δ^{18}O values were also highest in the early Holocene section of all the Agassiz cores. Glacio-isostatic evidence indicates deglaciation was underway by the beginning of the Holocene and maximum uplift occurred between 7000 and 8000 BP in many areas. Deglacial marine sediments in Clements Markham Inlet resemble those characteristic of temperate, not polar, tidewater glaciers, suggesting that climatic conditions in the early Holocene were significantly warmer than today (Stewart, 1988). Glaciers were behind modern terminal positions in some areas by 7500 ± 250 BP, and by 5700 ± 300 elsewhere. Marine mammals were present far beyond their normal modern range by 6500–7500 BP, as were many species of plants, between 9200 and 6700 BP In the northernmost valleys of Ellesmere Island and Peary Land, caribou were able to survive by (at least) 8500 ± 200 BP and driftwood was washing up on northern Ellesmere by 8800 BP Such evidence points to very warm conditions early in the Holocene (before 7500 ^{14}C BP, or 8000 calendar years) There is no doubt that the mid-Holocene was also quite warm, but evidence from maximum regional pollen rain and maximum driftwood occurrence may reflect *continuing* warm conditions, during which arctic vegetation became more widespread and driftwood dispersal from the more extensive northern forests became increasingly common. It is not possible to be certain about whether the early Holocene was warmer than the mid-Holocene, but there is certainly considerable evidence that this may have been the case. A consideration of solar radiation receipts due to changes in orbital configuration provides some support for this argument. Figure 13 shows June, July and August radiation anomalies (at the top of the atmosphere) for 80°N from 20,000 (calendar years) BP to the present Maximum anomalies of around +8% occurred between 11,000 and 8000 BP (although ^{14}C calibration with calendar years is not available before ~8000 ^{14}C BP, this period of radiation maximum probably corresponds to about 10,500 to 7000 ^{14}C years BP) It is worth noting that orbitally-induced radiation anomalies are especially significant at high latitudes in summer when daylight persists for 24 hours. Tricot and Berger (1988) calculate that the 10,000 BP July anomaly of radiation incident at the surface (i.e taking into account the effects of

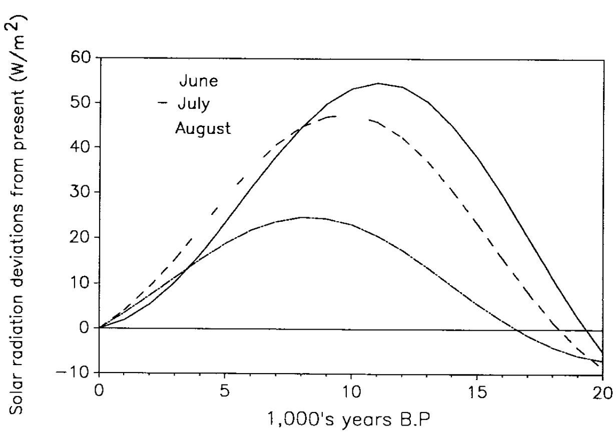

FIG 13. Daily radiation anomalies at the top of the atmosphere at 80°N in mid-June, mid-July and mid-August over the last 20,000 calendar years.

atmospheric depletion) was +20–25 W m^{-2} at 75–85°N This compares to modern (measured) values in the Queen Elizabeth Islands of 425 ± 25 W m^{-2} for mid-July (Maxwell, 1980) Such anomalies of incident radiation are higher than at any latitude, at any time, for at least 100,000 years. Of course, solar radiation receipts are not the only factor in raising summer temperature, advective effects are also very important. Nevertheless, the early Holocene radiation surplus must be considered as a significant factor, in view of the considerable evidence for exceptionally warm conditions at that time

Summer temperatures have declined significantly in the last 3500 ± 500 years

There is an abundance of evidence that summer climate has deteriorated over the last 3500 years. This includes a pronounced decline of 1.5–2.0‰ in $\delta^{18}O$ values in all Agassiz Ice Cap and Devon Island Ice Cap ice cores. The melt record from the Agassiz ice core indicates a decline in summer temperatures since ~5000 (^{14}C) BP (5500 calendar years) especially after ~2000 (^{14}C) BP Pollen and diatom records in lake sediments from southeastern Ellesmere Island indicate a decline in regional plant cover and increased summer ice cover on the lake. There may also have been a decline in peat growth since 3500 BP, although the evidence is equivocal Driftwood occurrence declined after 3500 BP and no driftwood younger than 3000 BP has been found behind the ice shelves of Ellesmere Island, suggesting that ice shelves formed after that time. Glaciers in many areas re-advanced between 3000 and 1000 BP, isolating proglacial lakes and over-riding organic material This climatic deterioration continued until the early part of this century, culminating in the exceptionally cold period of the last few centuries.

Climatic conditions were particularly severe from 100 to 400 BP

It is clear that the last few hundred years have been exceptionally cold. There is widespread evidence of glaciers reaching their maximum post-glacial positions within the last few hundred years, and the lowest $\delta^{18}O$ values and melt percentages for at least 1000 years are recorded in ice cores for this interval. On the Devon Island Ice Cap the period A.D 1550–1620 is considered to have been a time of net summer accumulation with very extensive summer sea-ice. A cold interval is also recorded by fine-grained ice on the Meighen Island Ice Cap, and regional pollen rain on the Agassiz Ice Cap reached the lowest levels for over 1000 years. Ice shelves probably reached their most extensive limits in the Holocene during this interval. Overall, the period from A.D 1550 to 1900 may have been the coldest period in the entire Holocene.

Summer temperatures since ~1925 have been exceptionally high compared to most of the late Holocene

Ice core records clearly show the exceptional nature of the last half century or so Melt percentages are extremely high, as are $\delta^{18}O$ values, and mass balance studies indicate many glaciers and ice caps have had a negative mass balance since ~1925 On both the Devon Island and Agassiz Ice Caps, melt percentages indicate that this period is the warmest for over 1000 years. Glaciers have retreated from many areas, revealing plants that were over-ridden during the glacial advances of the previous few centuries. Driftwood along the high water line is abundant in many parts of the archipelago and extensive break-up of ice shelves has been observed this century The last fifty years are quite exceptional in the context of the late Holocene, and perhaps comparable with warmer periods in the early to mid-Holocene.

SUMMARY

A wide variety of information relating to Holocene paleoclimatic conditions in the Queen Elizabeth Islands archipelago has been reviewed. Most proxy data reflect summer temperature conditions. Almost nothing can be said about winter paleoclimate Information has been derived from. ice cores, in particular changes in $\delta^{18}O$, ice fabric (especially the occurrence of features indicating melting and superimposed ice formation), cation concentrations and pollen content, lake sediments, including pollen, diatom and sedimentological changes, peat growth episodes and pollen content, driftwood frequency and the presence or absence of ice shelves, glacio-isostatic evidence, particularly the timing of deglaciation and of maximum uplift rates, glacial deposits and organic materials over-ridden by glacial advances, or exposed by ice recession, changes in the range of plant and animal species to localities beyond those of today

These diverse sources of paleoclimatic information indicate that summer temperatures in the early to mid-Holocene were similar to, or higher than, temperatures prevalent for much of this century when glaciers have been generally experiencing negative mass balance Summer temperatures declined after 3000 BP, reaching lowest levels from 100 to 400 BP Temperatures increased significantly after 1925 and may have been higher than at any time in the last 1000 years, and perhaps as high as any period of comparable length in the last 3000 years. Consequently the modern climate in the Queen Elizabeth Islands may be atypical of late Holocene conditions, but more characteristic of conditions which prevailed in the early to mid-Holocene.

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REFERENCES

Alt, B.T (1979). Investigation of summer synoptic climate controls on the mass balance of Meighen Ice Cap. *Atmosphere-Ocean*, **17** 181–199

Alt, B.T (1985). 1550–1620: a period of summer accumulation in the Queen Elizabeth Islands. *In*: Harington, C.R. (ed.), Climatic Change in Canada 5, *Syllogeus* **55**, pp. 461–480. National Museum of Canada, Ottawa.

Alt, B.T. *et al.* (1985). Arctic climate during the Franklin era, as deduced from ice cores. *In*: Sutherland, P.D. (ed.), The Franklin Era in Canadian Arctic History 1845–1859, pp. 69–92. National Museum of Man, Mercury Series, *Archaeological Survey of Canada, Paper No. 131* Ottawa.

Andrews, J T (1976) Little Ice Age permanent snowcover in the eastern Canadian Arctic: extent mapped from LANDSAT 1 satellite imagery *Geografiska Annaler* **58A (1–2)**, 71–81.

Barr, W (1971). Postglacial movement in northeastern Devon Island: a reappraisal. *Arctic*, **24**, 249–268.

Bednarski, J (1986) Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. *Canadian Journal of Earth Sciences*, **23**, 1343–1355.

Bednarski, J (1987). Glacial advances in Otto Fiord, Ellesmere Island, N W.T *Abstracts of the 16th Annual Arctic Workshop, Edmonton*, 4–5.

Bednarski, J (1988). The geomorphology of glaciomarine sediments in a high Arctic fiord. *Géographie Physique et Quaternaire*, **42**, 65–74.

Bednarski, J. (*in press*). An early Holocene bowhead whale (*Balaena myticetus* in Nansen Sound, Canadian Arctic archipelago. *Arctic*.

Bergsma, B.M. *et al.* (1984). Entombed plant communities released by a retreating glacier at central Ellesmere Island. *Arctic*, **37 (1)**, 49–52.

Bilello, M. (1966). Survey of Arctic and Subarctic temperature inversions. *Technical Report, 161* U.S. Army Materials Command, Cold Regions Research and Engineering Laboratory Hanover, N.H. 38 pp.

Blake, W Jr (1972). Climatic implications of radiocarbon-dated driftwood in the Queen Elizabeth Islands, Arctic Canada. *In*. Vasari, Y *et al.* eds), *Climatic Changes in Arctic Areas During the Last Ten Thousand Years*, pp. 77–104. University of Oulu, Oulu, Finland.

Blake, W Jr (1978). Coring of Holocene pond sediments at Cape Herschel, Ellesmere Island, arctic archipelago. *Current Research, Part C, Paper 78-1C*, pp. 199–122. Geological Survey of Canada, Ottawa.

Blake, W., Jr (1981a). Lake sediment coring along Smith Sound, Ellesmere Island and Greenland. *Current Research, Part A, Paper 81-1A* pp. 191–200. Geological Survey of Canada, Ottawa.

Blake, W. Jr (1981b). Neoglacial fluctuations of glaciers, southeastern Ellesmere Island, Canadian Arctic Archipelago. *Geografiska Annaler*, **63A**, 201–218.

Blake, W Jr (1982). Geological Survey of Canada Radiocarbon Dates XXII. *Paper 82-7 Geological Survey of Canada*, Ottawa, 22 pp.

Blake, W Jr (1984). Post-hypsithermal advance of Leffert Glacier Ellesmere Island, Arctic Canada. *Abstracts, 8th Biennial Meeting, AMQUA* 14.

Blake, W Jr (1985). Radiocarbon dating with accelerator mass spectrometry: results from Ellesmere Island, District of Franklin. *Current Research, Part B, Paper 85-1B*, pp. 423–429 Geological Survey of Canada, Ottawa.

Blake, W., Jr (1986). New AMS radiocarbon age determinations from east-central Ellesmere Island: applications to glacial geology *Current Research, Part B, Paper 86-1B*, pp. 239–244. Geological Survey of Canada, Ottawa.

Blake, W Jr (1987a). Lake sediments and glacial history in the High Arctic: evidence from east-central Ellesmere Island, Arctic Canada and from Inglefield Land, Greenland. *Polar Research*, **5**, 341–343.

Blake, W Jr (1987b). Geological Survey of Canada Radiocarbon Dates XXVI. *Paper 86-7* Geological Survey of Canada, Ottawa, 60 pp.

Blake, W Jr (1988). Geological Survey of Canada Radiocarbon Dates XXVII. *Paper 87-7* Geological Survey of Canada, Ottawa, 100 pp.

Blake, W Jr (1989). Inferences concerning climatic change from a deeply frozen lake on Rundfjeld, Ellesmere Island, Arctic Canada. *J Paleolimnology* **2**, 41–54.

Bliss, L.C. (1975). Devon Island, Canada. *In*: Rosswall, T and Head, O.W (eds), Structure and Function of Tundra Ecosystems. *Ecological Bulletin (Stockholm)* **20**, 17–60.

Bourgeois, J.C. (1986). A pollen record from the Agassiz Ice Cap, northern Ellesmere Island, Canada. *Boreas* **15**, 345–354.

Bourgeois, J.C. *et al.* (1985). 'Airborn pollen: a unique air mass tracer its flux to the Canadian High Arctic. *Annals of Glaciology* **7** 109–116.

Bovis, M.J and Barry R.G (1974). A climatological analysis of north polar desert areas. *In*: Smiley T.L. and Zumberge, J.H. (eds), *Polar Deserts and Modern Man*, pp. 23–31 University of Arizona Press.

Bradley R.S. and Eischeid, J.K. (1985). Aspects of the precipitation climatology of the Canadian High Arctic. *In*: Bradley, R.S. (ed.), *Glacial Geologic and Glacio-climatic Studies in the Canadian High Arctic*, pp. 250–271 Contribution No. 49, Department of Geology and Geography University of Massachusetts, Amherst.

Bradley R.S. and England, J (1978). Recent climatic fluctuations of the Canadian High Arctic and their significance for glaciology *Arctic and Alpine Research*, **10**, 715–731

Brassard, G.R. (1971). The mosses of northern Ellesmere Island, Arctic Canada. Ecology and phytogeography with an analysis for the Queen Elizabeth Islands. *The Bryologist*, **74**, 234–281

Brassard, G.R. and Blake, W Jr (1978). An extensive subfossil deposit of the arctic moss *Aplodon wormskioldii. Canadian Journal of Botany* **56**, 1852–1859.

Crary A.P (1956). Arctic Ice Island Research. *Advances in Geophysics* **III**, 1–41.

Crary A.P (1960). Arctic ice islands and ice shelf studies, Part III. *Arctic*, **13**, 32–50.

Crary A.P Kulp, J.L. and Marshall, E.W (1955). Evidences of climatic change from ice island studies. *Science* **122**, 1171–1173.

Dansgaard, W Johnsen, S.J Clausen, H.B. and Langway C.C. (1971). Climatic record revealed by the Camp Century ice core. *In*: Turekian, K.K. (ed.), *The Late Cenozoic Glacial Ages* pp. 37–56. Yale University Press.

Dansgaard, W Johnsen, S.J Claussen, H.B. and Gundestrup, N (1973). Stable isotope glaciology *Meddelelser om Gronland*, **197** 1–53.

Dansgaard, W White, J W.C. and Johnsen, S.J (1989). The abrupt termination of the Younger Dryas climate event. *Nature* **339**, 532–534.

Dyck, A. and Fyles, J.G (1963). Geological Survey of Canada Radiocarbon Dates II. *Radiocarbon*, **5**, 39–55.

Dyke, A.S. and Prest, V.K. (1987). Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. *Géographie Physique et Quaternaire*, **41**, 237–264.

Edlund, S.A. (1983). Bioclimatic zonation in a High Arctic region: central Queen Elizabeth Islands. *Current Research, Part A, Paper 83-1A*, pp. 381–390. Geological Survey of Canada, Ottawa.

Edlund, S.A. (1984). High Arctic plants: new limits emerge. *Geos*, **13**, 10–14.

Edlund, S.A. (1985). Lichen-free zones as neoglacial indicators on western Melville Island, District of Franklin. *Current Research, Part A, Paper 85-1A* pp. 709–712. Geological Survey of Canada, Ottawa.

Edlund, S.A. (1986). Modern Arctic vegetation and its congruence with summer climate patterns. *In*: French, H.M. (ed.), *Proceedings of Climatic Change on the Canadian Arctic*, pp. 84–99. Atmospheric Environment Service, Downsview Ontario.

Edlund, S.A. and Alt, B.T (1989). Regional congruence of vegetation and summer climate patterns in the Queen Elizabeth Islands NWT Canada. *Arctic*, **42**, 3–23.

Edlund, S.A. Alt, B.T and Young, K.L. (1989). Interaction of climate, vegetation and soil hydrology at Hot Weather Creek, Fosheim Peninsula, Ellesmere Island NWT *Current Research, Part D, Paper 89-1D* pp. 125–133. Geological Survey of Canada, Ottawa.

England, J (1978). The glacial geology of northeastern Ellesmere Island, N W T Canada. *Canadian Journal of Earth Science* **15**, 603–617

England, J (1982). Postglacial emergence along northern Nares Strait. *Meddelelser om Gronland*, **8**, 65–75.

England, J (1983). Isostatic adjustments in a full glacial sea. *Canadian Journal of Earth Sciences*, **20**, 895–917

England, J.H. (1987). *In situ* shells of full glacial age beneath former ice shelves. *Abstracts, 16th Annual Arctic Workshop, Edmonton*, pp. 20–22.

England, J.H. and Bednarski, J (1986). Postglacial isobases from northern Ellesmere Island and Greenland: new data. *Géographie Physique et Quaternaire*, **40**, 299–305.

England, J and Bradley R.S. (1978). Past glacial activity in the Canadian High Arctic. *Science*, **200**, 265–270.

Evans, D.J.A. (1988). *Glacial geomorphology and Late Quaternary history of Philips Inlet and the Wootton Peninsula, Northwest Ellesmere Island, Canada*. Ph.D Dissertation, University of Alberta, Edmonton, Canada, 281 pp.

Evans, D.J.A. (1989). An early Holocene narwhal tusk from the Canadian High Arctic. *Boreas* **18**, 43–50.

Fisher D.A. (1982) Carbon-14 production compared to oxygen isotope records from Camp Century Greenland and Devon Island, Canada. *Climatic Change* **4**, 419–426.

Fisher D.A. and Koerner, R.M. (1981 Some aspects of climatic change in the High Arctic during the Holocene as deduced from ice cores. *In*: Mahaney W.C. ed. *Quaternary Paleoclimate*, pp. 249–271 Geobooks, Norwich.

Fisher D.A. and Koerner R.M. (1983). Ice core study· a climatic link between the past, present and future. *In*: Harington, C.R. ed.), Climatic Changes in Canada 3, *Syllogeus* **49**, pp. 50–67 National Museum of Natural Sciences, Ottawa.

Fisher D.A. and Koerner R.M. (1988) The effects of wind on $\delta^{18}O$ and accumulation give an inferred record of seasonal δ amplitude from the Agassiz ice cap, Ellesmere Island, Canada. *Annals of Glaciology* **10**, 34–37

Fisher D.A. Koerner R.M. Paterson, W.S.B Dansgaard, W Grudenstrup, N and Reeh, N (1983). Effect of wind scouring on climatic records from ice-core oxygen-isotope profiles. *Nature* **301** 205–209

Häggblom, A. (1982). Driftwood in Svalbard as an indicator of sea ice conditions. *Geografiska Annaler* **64A**, 81–94.

Hammer C.U Clausen, H.B. and Tauber H. (1986). Ice core dating of the Pleistocene/Holocene boundary applied to a calibration of the ^{14}C time scale. *Radiocarbon*, **28**, 284–291

Harington, C.R. (1975). A postglacial walrus (*Odobenus rosmarus* from Bathurst Island, N W T *Canadian Field Naturalist*, **89**, 249–261.

Hattersley-Smith, G (1963). Climatic inferences from firn studies in northern Ellesmere Island. *Geografiska Annaler* **45**, 139–151

Hattersley-Smith, G (1969). Glacial features of Tanquary Fiord and adjoining areas of northern Ellesmere Island, N W T *Journal of Glaciology* **8 (52)**, 23–50.

Hattersley-Smith, G (1972). Climatic change and related problems in northern Ellesmere Island, N W T. Canada. *In*. Vasari, Y *et al.* (eds), *Climatic Changes in Arctic Areas During the Last Ten Thousand Years* pp. 137–148. University of Oulu, Oulu, Finland.

Hegg, O. (1963). Palynological studies of a peat deposit in front of the Thompson Glacier *Preliminary Report, 1961–1962, Axel Heiberg Island Research Report*, pp. 217–219 McGill University Montreal.

Hodgson, D.A. (1985). The last glaciation of west-central Ellesmere Island, Arctic Archipelago, Canada. *Canadian Journal of Earth Sciences* **22**, 347–368.

Holdsworth, G (1971). Calving from Ward Hunt Ice Shelf, 1961–1962. *Canadian Journal of Earth Sciences* **8**, 299–305.

Holdsworth, G (1977). Tidal interaction with ice shelves. *Annales de Géophysique* **33**, 133–146.

Hyvärinen, H. (1985). Holocene pollen stratigraphy of Baird Inlet, east-central Ellesmere Island, Arctic Canada. *Boreas* **14**, 19–32.

Jacobs, J.D and Leung, C.Y Y 1981) Paleoclimatic implications of topoclimatic diversity in Arctic Canada. *In*: Mahaney W.C. ed.), *Quaternary Paleoclimate*, pp. 63–75. GeoAbstracts, Norwich.

Jankovska, V and Bliss, L.C. (1977). Palynological analysis of a peat from Truelove Lowland. *In*: Bliss, L.C. (ed.) *Truelove Lowland, Devon Island, Canada: A High Arctic Ecosystem*, pp. 139–142. University of Alberta Press, Edmonton.

Jeffries, M.O (1987). The growth, structure and disintegration of Arctic ice shelves. *Polar Record*, **24**, 631–649.

Kelly M. and Funder S. (1974). The pollen stratigraphy of Late Quaternary lake sediments of south-west Greenland. *The Geological Survey of Greenland*, **64**, 1–26.

Khotinskiy N.A. (1984). Holocene vegetation history *In*: Velichko, A.A. (ed.), *Late Quaternary Environments of the Soviet Union*, pp. 179–200. University of Minnesota Press, Minneapolis.

King, L. (1981). Gletschergeschichtliche Arbeiten im Gebeit zwischen Oobloyah Bay N Ellesmere Island, N W T Kanada. *Heidelberger Geographische Arbeiten*, **69**, 233–267

King, L. (1983). Contribution to the glacial history of the Borup Fiord area, northern Ellesmere Island, N.W T Canada. *In*: Schroeder-Lanz, H. ed.), *Late and Post-glacial Oscillations of Glaciers* pp. 305–323. A.A. Balkema, Rotterdam.

Koerner R.M. (1968). Fabric analysis of a core from Meighen Ice Cap, N W T Canada. *Journal of Glaciology* **7** 421–430.

Koerner R.M. (1970). Some remarks on superimposition of ice on the Devon Island Ice Cap, N W T Canada. *Geografiska Annaler* **52A**, 57–67

Koerner R.M. (1977a). Distribution of microparticles in a 299 m core through the Devon Island Ice Cap, N W T Canada. *In*. *Isotopes and Impurities in Snow and Ice* pp. 371–376. IASH Publication 118.

Koerner R.M. (1977b). Devon Island Ice Cap: core stratigraphy and paleoclimate. *Science* **196**, 15–18.

Koerner R.M. (1980). The problem of lichen-free zones in Arctic Canada. *Arctic and Alpine Research*, **12**, 87–94.

Koerner R.M. (1983). Accumulation, ablation and oxygen isotope variations on the Queen Elizabeth Island Ice Caps, Canada. *Journal of Glaciology* **22**, 25–41.

Koerner R.M. (1984). Canadian Arctic Islands: glacier mass balance and global sea level. *In*: *Glaciers, Ice Sheets and Sea Level: Effects of a CO_2-induced Climatic Change*, pp. 145–154. U.S. Department of Energy Washington D.C.

Koerner R.M. and Fisher D.A. (1985) The Devon Island ice core and the glacial record. *In*: Andrews, J T (ed. *Quaternary Environments: Eastern Canadian Arctic, Baffin Bay and Western Greenland*, pp. 309–327 G Allen & Unwin, Boston.

Koerner R.M. and Fisher, D.A. (1989). A 10,000 year record of summer climate: Agassiz Ice Cap, northern Ellesmere Island, Northwest Territories, Canada. *Nature (in press)*.

Koerner, R.M. and Paterson, W.S.B. (1974). Analysis of a core through the Meighen Ice Cap, Arctic Canada, and its paleoclimatic implications. *Quaternary Research*, **4**, 253–263.

Koerner R.M. and Russell, R.P (1979) $\delta^{18}O$ variations in snow on the Devon Island Ice Cap, N W T Canada. *Canadian Journal of Earth Sciences* **16**, 1419–1427

Koerner R.M. Paterson, W.S.B. and Krouse, H.R. (1973) A $\delta^{18}O$ profile in ice formed between the equilibrium and firn lines. *Nature* **243**, 137–140.

Koerner R.M. Bourgeois, J.C. and Fisher D.A. 1988). Pollen analysis and discussion of time-scales in Canadian ice cores. *Annals of Glaciology* **10**, 85–91

Leech, R.E. 1966). The spiders (*Areneida* of Hazen Camp, 81°49′N 71°18′W *Quaestiones Entomologicae* **2**, 153–212.

Lemmen, D.S. 1988). *The Glacial History of Marvin Peninsula, Northern Ellesmere Island and Ward Hunt Island, High Arctic Canada*. Ph.D Dissertation, University of Alberta, Edmonton, 176 pp.

Locke, C.W and Locke, W W 1977). Little Ice Age snow-core extent and paleoglaciation thresholds: north-central Baffin Island, N W T Canada. *Arctic and Alpine Research* **9**, 291–300.

Long, A. 1967) Age of trapped sea-water at bottom of Lake Tuborg, Ellesmere Island, N W T *Transactions of American Geophysical Union*, **48**, 136.

Lowdon, J.A. and Blake, W Jr 1979). Geological Survey of Canada Radiocarbon Dates XIX. *Paper 79-7* Geological Survey of Canada, Ottawa, 58 pp.

Lyons, J.B. and Mielke, J.E. (1973). Holocene history of a portion of northernmost Ellesmere Island. *Arctic* **26**, 314–323.

Lyons, J.B. and Ragle, R.H. (1962). Thermal history and growth of the Ward Hunt Ice Shelf. *UIGG-IAHS Commission des Neiges et Glace, Colloque d'Obergurgl*, pp. 88–97

Maxwell, J.B. (1980). The climate of the Canadian Arctic Islands and adjacent waters, Volume I. *Climatological Studies No. 30*. Atmospheric Environment Services, Downsview Ontario, 532 pp.

Maxwell, J.B (1981). Climatic regions of the Canadian Arctic Islands. *Arctic*, **34**, 225–240.

Maxwell, J.B. (1982). The climate of the Canadian Arctic Islands and adjacent waters, Volume II. *Climatological Studies No. 30*. Atmospheric Environment Services, Downsview Ontario, 589 pp.

Miller G.H. (1976). Anomalous local glacier activity Baffin Island, Canada. paleoclimatic implications. *Geology*, **4**, 502–504.

Miller G.H. Bradley R.S. and Andrews, J T (1975). The glaciation level and lowest equilibrium line altitude in the High Canadian Arctic: maps and climatic interpretation. *Arctic and Alpine Research*, **7** 155–168.

Müller, F (1964) Evidence of climatic fluctuations on Axel Heiberg Island, Canadian Arctic Archipelago. *In*: Fletcher J.O. (ed.), *Proceedings of the Symposium on the Arctic Heat Budget and Atmospheric Circulation*, pp. 136–156. Rand Corporation, Santa Monica, California.

Nichols, H. (1975). Palynological and paleoclimatic study of the Late Quaternary displacement of the boreal forest-tundra ecotone in Keewatin and MacKenzie, N W T Canada. *Occasional Paper No. 15* Institute of Arctic and Alpine Research, University of Colorado, Boulder

Ovenden, L. (1988). Holocene proxy-climate data from the Canadian Arctic. *Paper 88-22* Geological Survey of Canada, Ottawa.

Paterson, W.S.B (1968). A temperature profile through the Meighen Ice Cap, Arctic Canada. *International Association of Scientific Hydrology* Publication **79**, 440–449.

Paterson, W.S.B. (1969). The Meighen Ice Cap, Arctic Canada: accumulation, ablation and flow *Journal of Glaciology* **8**, 341–352.

Paterson, W.S.B. and Waddington, E.D (1984). Past precipitation rates derived from ice core measurement: methods and data analysis. *Reviews of Geophysics and Space Physics*, **22**, 123–130.

Paterson, W.S.B. Koerner R.M. Fisher, R.D. Johnsen, S.I. Clausen, H.B Dansgaard, W Butcher, P and Oeschger H. (1977) An oxygen isotope climatic record from the Devon Island Ice Cap, Arctic Canada. *Nature* **266**, 508–511

Retelle, M.J (1986a). Stratigraphy and sedimentology of high arctic coastal lacustrine basins, northeastern Ellesmere Island, Northwest Territories. *Géographie Physique et Quaternaire*, **40**, 117–128.

Retelle, M.J (1986b). Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, Northwest Territories. *Canadian Journal of Earth Sciences*, **23**, 1001–1012.

Retelle, M.J Bradley R.S. and Stuckenrath, R. (1989). Relative sea level chronology determined from raised marine sediments and coasted isolation basins, northeastern Ellesmere Island, Arctic Canada. *Arctic and Alpine Research*, **21**, 113–125.

Ritchie, J.C. (1984). *Past and Present Vegetation of the Far Northwest of Canada*. University of Toronto Press, Toronto, 251 pp.

Ritchie, J.C. and Hare, F.K. (1971). Late Quaternary vegetation and climate near the Arctic tree line of northwestern North America. *Quaternary Research*, **1**, 331–342.

Smol, J.P (1983). Paleophycology of a high arctic lake near Cape Herschel, Ellesmere Island. *Canadian Journal of Botany* **61**, 2195–2205.

Smol, J.P (1988). Paleoclimate proxy data from freshwater arctic diatoms. *Verh. Int. Verein. Limnology* **23**, 837–844.

Stewart, T.G (1988). *Deglacial-Marine Sediments from Clements Markham Inlet, Ellesmere Island, N.W T., Canada*. Ph.D Dissertation, University of Alberta, Edmonton, Canada, 229 pp.

Stewart, T.G and England, J (1983). Holocene sea-ice variations and paleoenvironmental change, northernmost Ellesmere Island, N W T Canada. *Arctic and Alpine Research*, **15**, 1–17

Stewart, T.G and England, J (1986). An early Holocene caribou antler from northern Ellesmere Island, Canadian Arctic archipelago. *Boreas* **15**, 25–31

Stuiver M. (1982). A high precision calibration of the A.D radiocarbon time-scale. *Radiocarbon*, **24**, 1–26.

Völk, H.R. (1980). Record of emergence around Oobloyah Bay and Neil Peninsula in connection with the Wisconsin deglaciation pattern, Ellesmere Island, N W T Canada. a preliminary report. *Polarforschung* **50**, 29–44.

Williams, L.D and Bradley R.S. (1985). Paleoclimatology of the Baffin Bay Region. *In*: Andrews, J T (ed.), *Quaternary Environments. Eastern Canadian Arctic, Baffin Bay and Western Greenland*, pp. 741–772. Allen and Unwin, Boston.