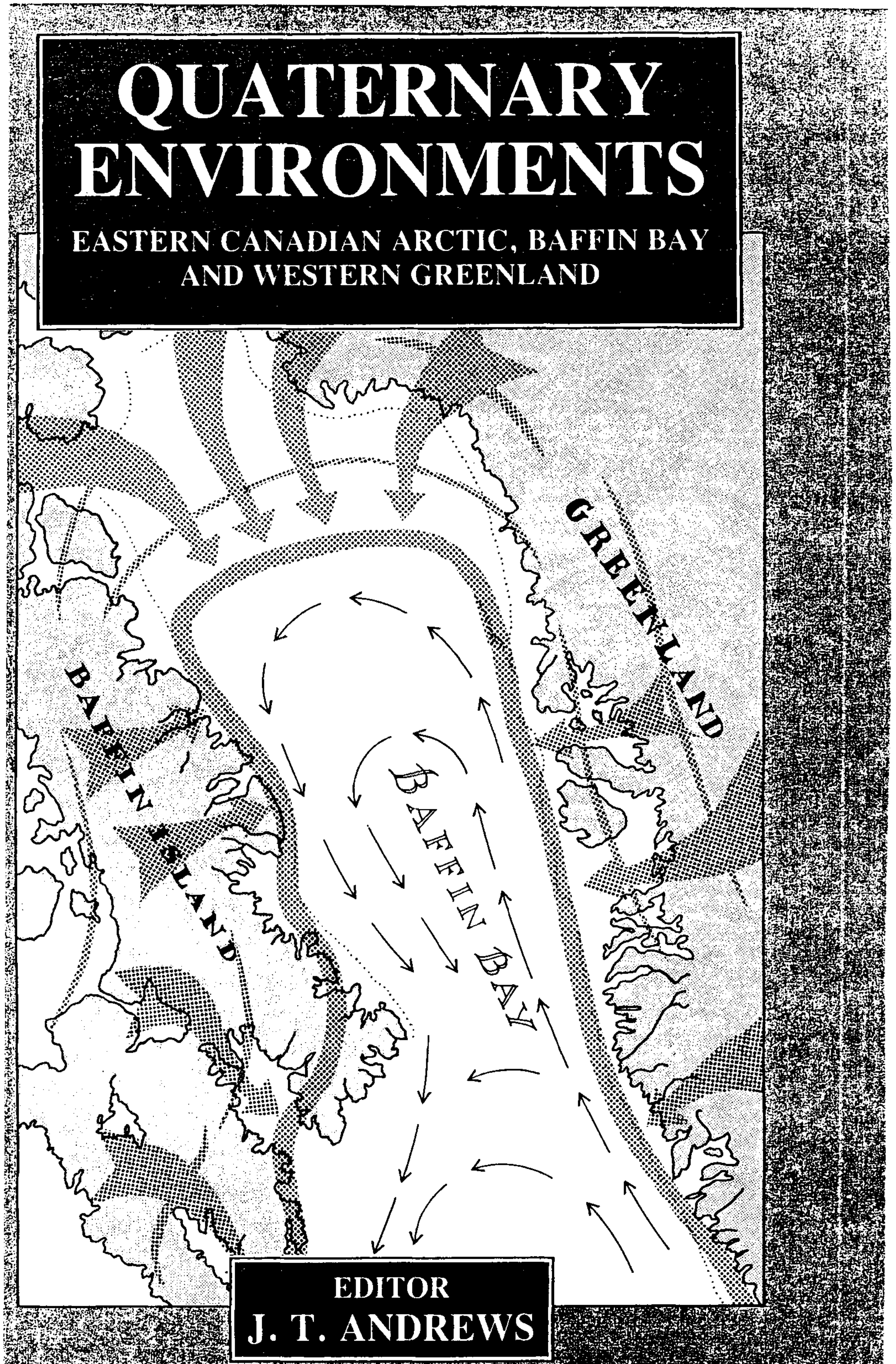


26 Paleoclimatology of the Baffin Bay region

L. D. Williams and R. S. Bradley

The paleoclimatology of the Baffin Bay region is based on a variety of proxy data, some of which have been discussed in considerable detail in previous chapters. Here, we compare the paleoclimatic record from various sources and attempt to identify areas of agreement as well as inconsistencies in the record, as presently understood. First, however, we consider the value of paleoclimatic studies in this region.

THE SIGNIFICANCE OF BAFFIN BAY REGION PALEOCLIMATE

How important are climatic fluctuations of the Baffin Bay area (= Baffin region)? What significance do Baffin region paleoclimatic data have for other Arctic areas? Meteorological records, in the Arctic are, unfortunately, quite brief (Chapter 2) but studies of data from recent decades clearly indicate that temperature fluctuations in the Baffin region are highly correlated with those elsewhere in the Arctic. For example, Walsh (1977) showed that in all seasons, the principal eigenvectors of Arctic surface air temperatures were centered over Baffin Island, indicating temperature variations in the Baffin region are typical of a very large part of the Arctic. Similarly, Keen (1980) demonstrated that temperature fluctuations (1951-76) at c. 80°W are more highly correlated with the 70°N zone average than any other zone in the Arctic (Fig. 26.1). Consequently, Keen concluded that "Baffin temperature [is] a sensitive indicator of summer conditions across the Arctic as a whole." If such relationships are typical of longer time scales (probably a reasonable assumption for periods when surface boundary conditions were not drastically different from those of recent decades) then the paleoclimatology of the Baffin region assumes added importance. Paleoclimatic fluctuations of the Baffin region are thus of significance to studies of the entire Arctic.

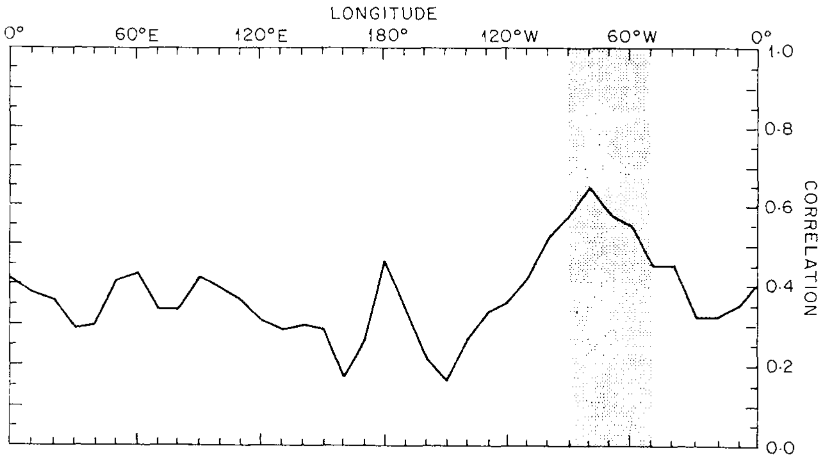


Figure 26.1 Correlation coefficient of summer (J,J,A) mean temperature at different longitudes at 70°N with the zonal average (1951-76). Shaded area signifies longitudinal zone from West Greenland to the west coast of Baffin Island at 70°N. Temperatures in this zone are highly correlated with the zonal mean (after Keen, 1980) (with permission).

It has often been said that the Baffin region is particularly 'sensitive' to climatic fluctuations (e.g. Tarr, 1897; Andrews et al., 1972). This sensitivity results from two main factors: low summer temperatures and the position of Baffin Island in relation to the mean position of the mid-tropospheric trough over northeastern North America. On a latitudinal basis (Fig. 26.2) summer temperatures on Baffin Island are lower than in any other sector of the Arctic (except for the Greenland Ice Cap) (Barry et al., 1977; Keen, 1980). Thus, relatively small changes in summer temperature markedly affect snow cover conditions and glacier mass balance (Bradley and Miller, 1972). In particular, the upland regions of Baffin Island, which are extensive, are near the threshold of glacierization; relatively small decreases in ablation season temperature would initiate large scale expansion of firn on these uplands (Williams, 1978). Indeed, there is much evidence that this did occur during the Little Ice Age (17th-19th centuries) (Locke and Locke 1977; Andrews et al., 1976).

The sensitivity of Baffin Island is also closely related to the regional circulation patterns. A major feature of the northern hemispheric general circulation is an upper level trough over Baffin Bay. The precise position of this trough is very important for large-scale advection of air across Baffin Island. If the trough is displaced westward, relatively warm southerly and southeasterly winds

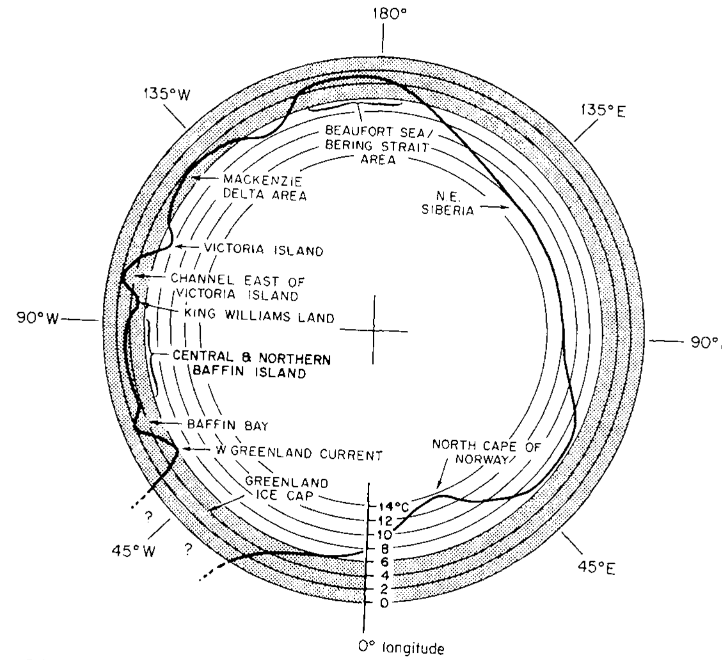
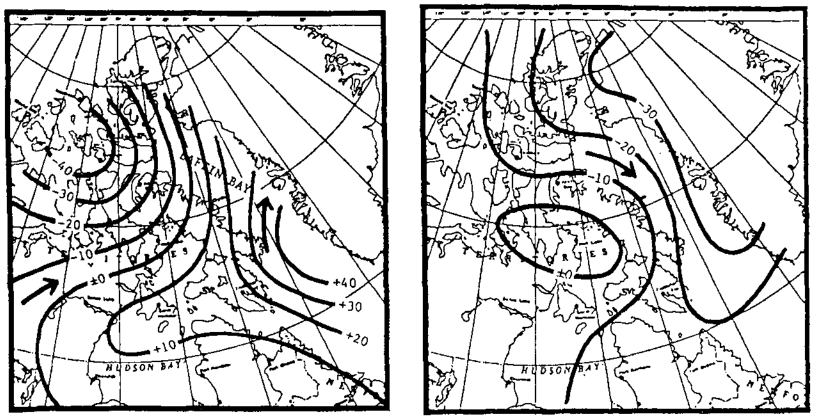


Figure 26.2 Mean July temperature at 70°N around the Northern Hemisphere (from Barry et al., 1975) (with permission).

increase in frequency leading to above average summer temperatures. Conversely, with an eastward displacement of the trough axis, airflow from the north and northwest increases in frequency resulting in colder, generally drier summer conditions (Fig. 26.3). Statistically, the relationship between mean trough position in summer and mean summer temperatures in the Baffin region is strong. For example, for the period 1949-1976, the correlation coefficient (r) was 0.57 (significant at >99% level) (Keen, 1980). Such changes in trough position should not be viewed in isolation from changes in the general circulation of which they are a part. In a study of the relationships between hemispheric circulation indices and summer temperatures in the Baffin area, Keen (1980) convincingly demonstrated that temperature changes on a hemispheric scale are accentuated in the Baffin region due to the effects of trough displacement (Fig. 26.4). As hemispheric cooling occurs, meridional temperature gradients are enhanced, leading to stronger mid-latitude westerly airflow. This, in turn, causes the wavelength of upper level waves in the atmosphere to increase, with the result that the Baffin trough is displaced eastward. The trough may also be deepened by increased cyclogenesis along the arctic front



(a) 1943-47 to 1951-60 (b) 1951-60 to 1964-73

Figure 26.3 Changes in mean July 700 mb heights durng two contrasting periods. From 1943-47 to 1951-60 the mean trough position migrated westward, resulting in increased southerly and westerly airflow over Baffin Island and consequently higher summer temperatures. By contrast, the mean trough position 1964-73 was further east, leading to enhanced northerly and northeasterly airflow and significantly lower summer temperatures (after Keen, 1980).

SUMMER HEMISPHERIC COOLING

↓

INCREASED MERIDIONAL TEMPERATURE GRADIENTS

↓

STRONGER MID-LATITUDE WESTERLY AIRFLOW

↓

INCREASE IN WAVELENGTH OF UPPER LEVEL ROSSBY WAVES

↓

TROUGH IN BAFFIN BAY DISPLACED EASTWARD

↓

NORTHERLY AIRFLOW INCREASED: SIGNIFICANTLY LOWER TEMPERATURES OVER BAFFIN ISLAND

Figure 26.4 Schematic diagram illustrating relationship between hemispheric cooling and enhanced temperature decline in the Baffin Island area.

leading to more depressions tracking northward into Baffin Bay. As a result, cold northerly airflow is drawn over Baffin Island, accentuating the hemispheric coolong in progress, and leading to reduced regional ablation and enhanced glacierization.

REGIONAL CLIMATIC GRADIENTS

Climatic conditions across Baffin Bay are by no means uniform (Figs 26.5, 26.6 & 26.7), and these sub-regional scale differences have significance for paleoclimatic studies. Of particular importance is the marked contrast in mean annual temperature on opposite sides of Baffin Bay which is related to oceanic circulation patterns. In eastern Baffin Bay/Davis Strait, the West Greenland Current carries relatively warm water (and associated warm air) along the west coast of Greenland. In western Baffin Bay, southward flowing, cold Arctic water prevails. As a result, although latitudinal temperature gradients are similar (a decrease of 0.65°C per 1° increase in latitude), mean annual temperatures on the west Greenland coast are 8°C warmer than at the same latitude on the east coast of Baffin Island (Fig. 26.5). South of

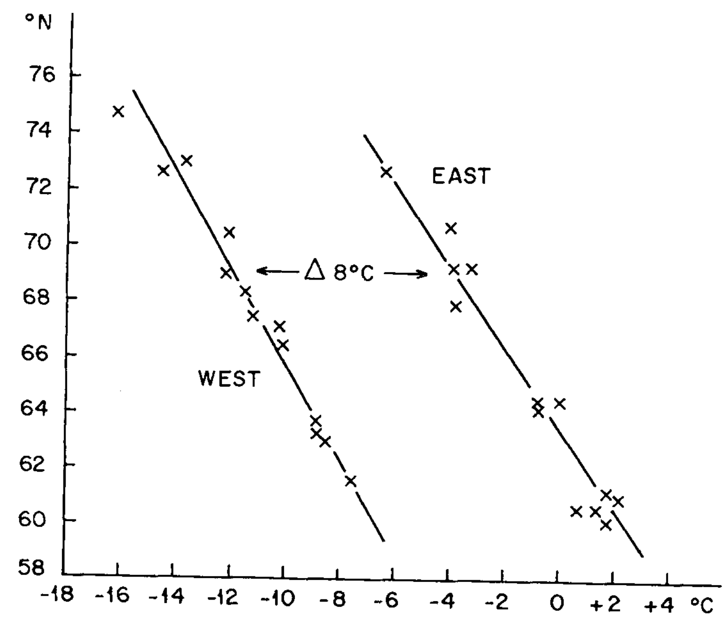


Figure 26.5 Latitudinal mean summer temperature gradients along the western and eastern sides of Baffin Bay and Davis Strait.

64°N, along the Greenland coast, open water is present throughout the year (Fig. 26.6(a)). However, on the western side of Baffin Bay,

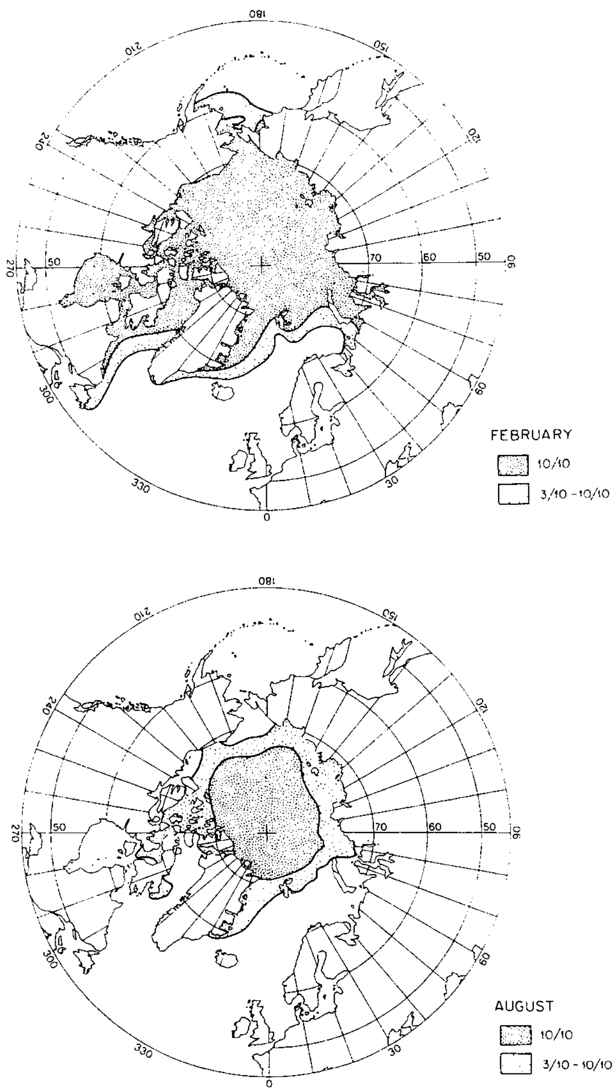


Figure 26.6 (a) Sea-ice distribution in February (maximum extent). (b) Sea-ice distribution in August (minimum extent). Note contrast in sea-ice extent in mid-winter between the western and eastern sides of Davis Strait/Baffin Bay.

mid-winter ice cover extends southward along the Labrador coast as far as 49°N and even in late summer, considerable ice may remain around the Cumberland Peninsula and within Foxe Basin (Fig. 26.6(b)). Sea-ice extent plays an important role in determining the amount of precipitation which occurs in different parts of the region. As shown in Figure 26.7, precipitation amounts increase significantly in southernmost Greenland (south of c. 62°N) where open water occurs throughout the year. On southeastern Baffin Island, heaviest precipitation amounts occur where locally extensive open water occurs year round, as, for example, near Cape Dyer (Fig. 26.6(a) & (b)). Elsewhere, mean annual precipitation amounts are generally les than 400 mm.

From this consideration of contemporary climatic conditions, it is apparent that oceanographic conditions play a very important role in regional climatic differences. More extensive open water in the past would have been associated with high temperature (though not necessarily in the ablation season). At the same time, there would

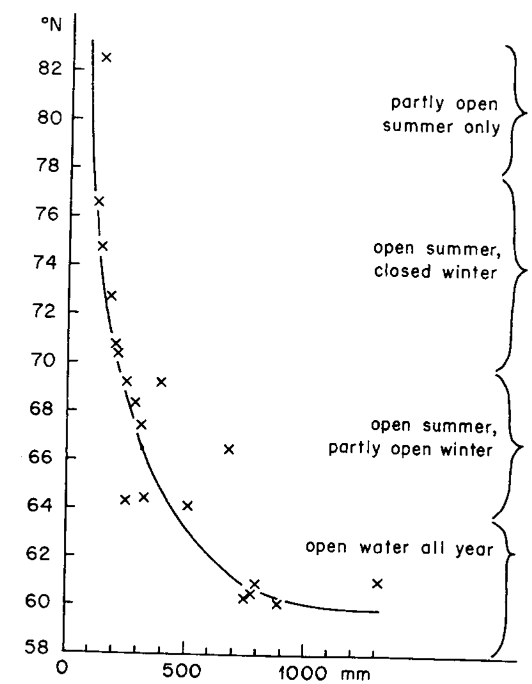


Figure 26.7 Mean annual precipitation amounts around Baffin Bay/Davis Strait, in relation to seasonal sea-ice extent. Note marked increase in precipitation towards zone with open water all year (southern Greenland). Locations with locally persistent polynas year round (e.g. Cape Dyer at c. 66°N) have anomalously heavy precipitation.

have been an even more pronounced increase in precipitation (since the precipitation/proximity to open water relationship is non-linear). Conversely, when the West Greenland current was absent due to more extensive Arctic water, as for example from >20,000 to c. 10,000 BP (Ruddiman and McIntyre, 1981), precipitation would have been greatly reduced and the temperature decrease along the West Greenland coast would have been much greater than on the coast of eastern Baffin Island. In short, strong differential changes in climate across Baffin Bay are likely to have occurred in the past as a result of changes in large-scale oceanographic conditions.

EVIDENCE USED IN PALEOCLIMATIC STUDIES

The principal lines of evidence used in the reconstruction of paleoclimates in the Baffin region pertain to temperatures during the summer, which is the environmentally sensitive period; large changes in winter temperatures may have taken place in the past, but there are few means of detecting if such variations occurred. It is possible that studies of periglacial features (e.g. ice and sand wedges) could provide a valuable winter season perspective on paleoclimatic conditions (cf. Hopkins, 1982).

Proxy data (Table 26.1) from the Baffin area are, of course, subject to the same limitations as similar data from other areas. The data are often of dubious paleoclimatic significance and may be discontinuous in time and space, providing only episodic information. On Baffin Island, only peat and lake sediments have provided continuous paleoclimatic records and pollen studies of these sediments have formed the basis of several regional quantitative paleoclimatic studies (e.g. Andrews et al., 1980; Andrews and Diaz, 1981; Diaz and Andrews, 1982). Not all of the records used in these studies are well dated; in particular, many lake sediment records from northern Labrador have dates which are out of sequence stratigraphically (Short and Nichols, 1977; Short, 1978), possibly as a result of disturbance and/or contamination during the coring operation (removal of lower core sections). Further lake sediment studies are required to resolve the many uncertainties in these records.

THE PALEOCLIMATIC RECORD

In this section we compare the various proxy data records which bear on paleoclimatic conditions in the Baffin Island region. Particular attention is given to events over a few decades (for the more recent

TABLE 26.1. Types of evidence used in paleoclimatic reconstructions in the Baffin Bay area.

Proxy Data	Continuous (C) or Discon-tinuous (D)	Method of Dating	Time Frame (year)	Dating Uncer-tainty Factor	Minimum Sampling Interval (yrs)*	Resolution (period length in years)	Paleo-climatic Inference	Remarks	Reference Example
Glacial Deposits (moraines, till, outwash)	D	C-14 lichens	0-10,000+ 5-5,000	±50-200 yrs. ±10%			(Advances): cooler summers and/or higher accumulation. (Recession): warmer summers and/ or less accumulation.	Glacier advance or retreat reflects mass balance changes; though these may be complex, they are generally ascribed to changes in summer temperatures.	Carrara and Andrews, 1972. Miller, 1973. Davis, this volume.
Soils	D	C-14	0-5,000+	±50-200			Colder after date on top surface of soil.	Dates generally on uppermost soil profile when overlain by sand, outwash or till. May be dating problem (J.A. Mathews, 1980).	Miller, 1973.
Peat Growth	C or D	C-14	0-7,000+	±50-200	50-600		Summers warmer during peat growth inter-val, cooler after peat growth ceases.	Basal and/or uppermost dates generally supple-mented by pollen data.	Nichols, 1974
Pollen in: a) lake sediment	C	C-14	0-9,200	±100-300		100-1200	Summers warmth; southerly airflow	Calibration with modern data possible to give quantitative	Davis, 1980. Short and Andrews, 1980.

b) peat	C or D	C-14	0-7,000	±50-7,000			frequency.	paleoclimatic estimates. Isotopic studies (O-18) possible also (e.g. Gray 1982).	Andrews et al. 1979, 1980.
Ice: a) O-18	C	Multivariate	0-10,000	±5% to 5,000, increasing to ±10% to 10,000	1-50+ increasing with time	2-100+	Snowfall event temperatures and/or proximity to open water and/or ice surface height.	Seasonal signals possible.	Koerner, 1977. Short and Andrews, 1980. Andrews et al. 1979 - 1980.
b) Melt Layers	C	Multivariate	0-2,000	±5%	1-10+	2-20	Summer temperature.	Percolation may cause meltwater to penetrate deeper layers, thereby "extending" period of apparent warmth.	Fisher and Koerner 1981. Herron et al. 1981.
Tree Rings	C	Annual ring counts	0-300	±1 yr.	1	2	Summer temperature.	Northern Labrador treeline only.	Cropper and Fritts, 1981.
Drift Wood	D	C-14	0-9,000+	±50-200 yrs.			Summer temperature.	Periods when driftwood frequency is highest are thought to reflect reduced sea-ice cover/open water.	Stewart and England, 1983.

records) to a few centuries. Of course, dating uncertainty and differing responses of different proxy climatic indicators make comparisons on these time scales difficult. However, some knowledge of short-term paleoclimatic variations would be useful, for comparison with similar changes in other regions, and with the short-term fluctuations of atmospheric C-14 (revealed by the departure of radiocarbon ages from dendrochronological ages), which may have been caused by solar variations (Denton and Karlen, 1973).

In view of the scarcity of information and dating uncertainty prior to the postglacial period, consideration is confined to the last 12,000 C-14 years. This period subdivides naturally, on the basis of records of comparable detail, into three nested intervals: (1) the last 12,000 C-14 years BP*, (2) 4000 BC to present, and (3) AD 600 to present. It many not be possible to identify short-term climatic changes in the earlier part of this period, the glacial to interglacial transition, for the evidence must reflect such things as changes in oceanic circulation, rates of glacier calving, and floral/faunal succession, as well as climatic variations. Even later evidence does not necessarily permit a simple climatic interpretion. Changes in vegetation, as measured by pollen assemblages, are probably indicative of summer temperature changes, although moisture conditions may also have been important (Andrews et al., 1980). In this regard, it should be noted that temperature may also control surface-water availability, at least locally, by its effect on thawing of frozen ground and the depth of the active layer in summer. Glacier fluctuations may be related to summer temperature, but could also be caused by changes in snowfall or even by non-climatic factors. The amount of refrozen meltwater in ice cores (Fisher and Koerner, 1981; Herron et al, 1981) can be directly attributed to summer temperature changes, although even this may be complicated by local factors such as radiation climate.

In general, we can only assume that most of the proxy climatic evidence considered (at least in the two later intervals) are indicative of summer temperature variations, with the understanding that some of the discrepancies observed may be due to violations of the various assumptions. Oxygen isotope variations in ice cores are problematical as whatever climatic information they contain is related to conditions on precipitation days, most of which occur in the months of May and September to November (Bradley, 1983). However, two records

* 'BP' is to be understood to mean conventional C-14 years before present.

of 0-18 maxima (i.e. maxima in the annual cycles) are included from cores for which data on intra-annual variations are available, and also two mean annual 0-18 records for comparison with meltwater records from the same sites.

The locations from which the data were obtained are shown in Figure 26.8. Some information from locations in the Canadian Arctic outside the Baffin Bay region is included in this study for comparison.

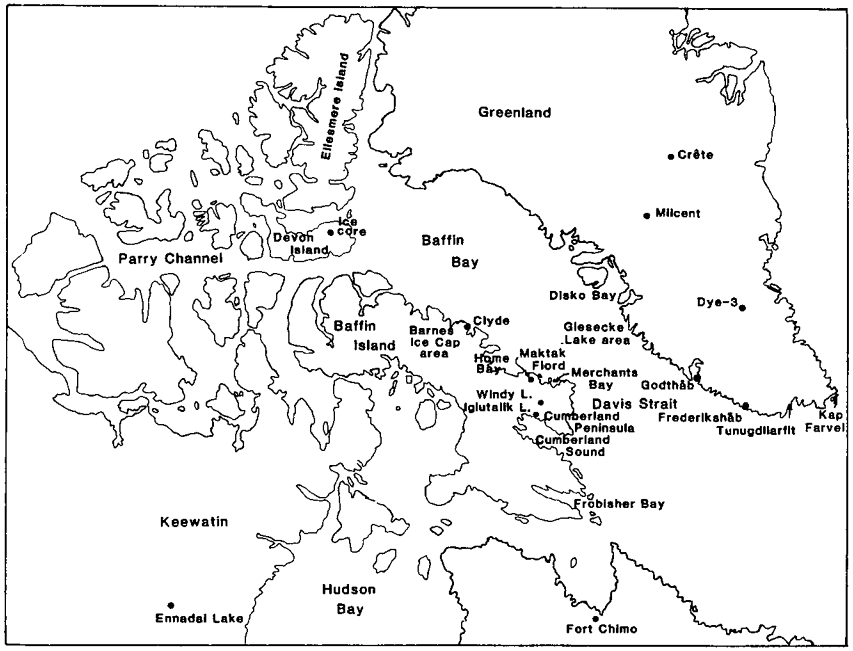


Figure 26.8 Location of places from which data have been used in this study.

THE LAST 12,000 C-14 YEARS

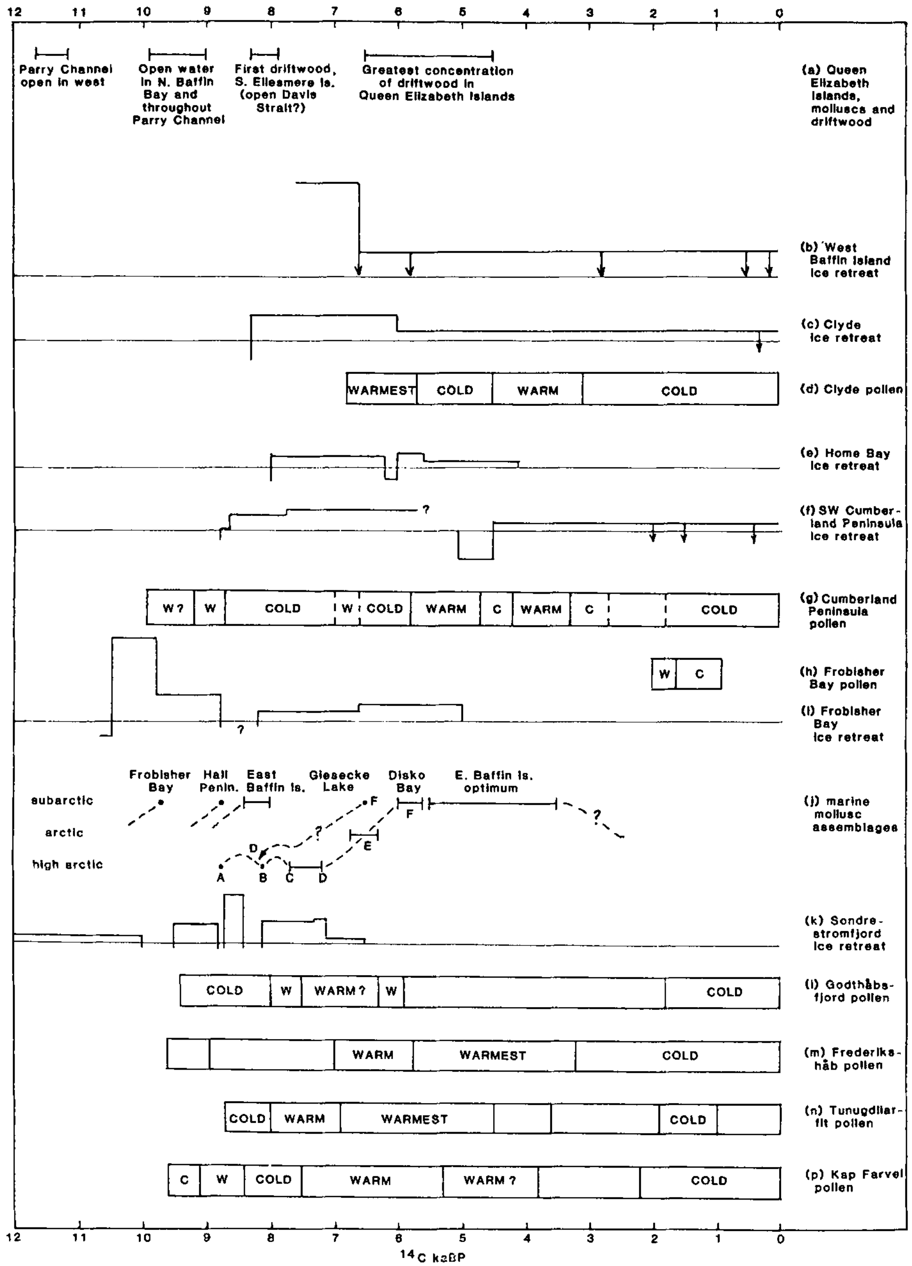
Figure 26.9 compares events during the last 12,000 C-14 years, arranged roughly from northwest at the top to southeast at the bottom. At the beginning of this period, it appears that there was a great difference in climatic conditions between the Baffin Bay region and the western Canadian Arctic. The first postglacial appearance of open water in western Parry Channel about 11,500 BP (Fig. 26.9a) coincided with the

advent of spruce on the Mackenzie River delta, which is north of the present tree line (Ritchie and Hare, 1971; Ritchie et al., 1983). The northwestern margin of the Laurentide ice sheet may have been retreating at the rate of about 200 m per year between 12,000 and 10,000 BP (Bryson et al., 1969; Prest, 1970). In contrast, the retreat rate of part of the ice sheet in southwestern Greenland during this time was only 10 m per year (Fig. 26.9k) and the ice margin in eastern Baffin Island appears to have been stable (Andrews, 1975).

By 10,000-9000 BP, there was open water throughout Parry Channel and in northern Baffin Bay (Fig. 26.9a). The correspondence between radiocarbon dates and actual (calendar) dates is not definitely known before 7000 BP, but at that time the C-14 dates are about 1000 years too young, and if such is also true of dates 10,000-9000 BP, then the advent of open water in northern Baffin Bay roughly coincided with the very abrupt rise of 0-18 values in the Camp Century ice core (Hammer et al., 1978). Meanwhile, in the southern part of the Baffin Bay region, there seems to have been a rapid retreat of the ice margin in Frobisher Bay (Fig. 26.9i) between 10,500 and 8800 BP (and also after c. 10,300 BP in Merchants Bay, not shown in Fig. 26.9). This may have been due partly to increased rates of calving as relative sea level rose (cf. Fig. 2 in Andrews, 1982), but it may also have been partly climatic, as suggested by exotic pollen influx on southwestern Cumberland Peninsula, 10,000-8700 BP (Fig. 26.9g; see Short et al., this volume), and the intrusion of subarctic species of molluscs in Frobisher Bay about 9700 BP (Fig. 26.9j).

In contrast, there was an advance or stillstand of the ice margin in southwestern Greenland between 10,000 and 9500 BP (Fig. 26.9k). However, another period of advance of the southwestern Greenland ice margin between 8800 and 8100 BP (although interrupted by a rapid retreat) does coincide with an extensive advance of the Laurentide ice margin on Baffin Island (the 'moraines of Cockburn age'; Andrews and

Figure 26.9 Some events of the last 12,000 C-14 years in the region. (a) Inferences made by Blake (1972) on the basis of fossil molluscs and driftwood in the Queen Elizabeth Island. (b,c,e,f,i,k) Relative rates of ice margin retreat (up) or advance (down) estimated on Baffin Island by Andrews (1982) and on southwestern Greenland by Ten Brink and Weidick (1974). (d,g,h,k,l,m,n,p) Relative changes in summer temperature inferred from pollen assemblages on Baffin Island (Short et al., this volume) and in southwestern Greenland (Fredskild, this volume). (j) Changes in mollusc assemblages on the coasts of Baffin Island (Andrews, 1972; Miller, 1980) and West Greenland (Laursen, 1976).



Ives, 1978). Pollen assemblages from Godthaabsfjord and Tunugliarfik (Figs 26.91 & n) indicate that this was a relatively cold period. The pollen record from Kap Farvel (Fig. 26.9p) disagrees in part, but would correlate well with the other evidence if its time scale was a few centuries older.

The glacial records from Clyde, Home Bay, Frobisher Bay, and Sondre Stromfjord (Figs 26.9c, e, i, & k) all agree on the termination date of this glacial stage at 8300-8000 BP. This coincides with the first appearance of driftwood on southern Ellesmere Island (Fig. 26.9a) (which suggested to Blake, 1972 the opening of Davis Strait) and with the (probably related) intrusion of subarctic species of molluscs on the east coast of Baffin Island (Fig. 26.9j). It also coincides with the beginning of a long period of warmth-indicating pollen assemblages in southwestern Greenland (again, with Kap Farvel slightly later).

The evidence from Cumberland Sound on glacier retreat (Fig. 26.9f) and pollen (Fig. 26.9g) appears to be contradictory, both with each other and with the other records. There, ice margin retreat seems to have commenced with the onset of a cold period about 8700 BP which lasted through most of the following three millenia. Another apparent contradiction to climatic amelioration about 8000 BP is the continued presence of high arctic mollusc species in the Disko Bay (West Greenland) area until about 7000 BP (horizons C and D, Fig. 26.9j). This may be a matter of succession rather than climate as both Laursen (1976) and Miller (1980) have noted the time-transgressive spread of subarctic molluscs on the coasts of West Greenland and Baffin Island, respectively (see Fig. 26.9j; see Fig. 26.8 for the locations). Curiously, the subarctic species seem to have arrived first on the Baffin Island coasts, although today they are restricted in this region to the warm West Greenland Current (Andrews, 1972).

Andrews and Ives (1972) suggested that ice margin advance (or stillstand) on Baffin Island in the Cockburn Stade might be attributed to increased snowfall due to the incursion of warmer water into Davis Strait while the land remained cold. Given the dating uncertainty, this is certainly a possibility, but an alternative explanation could be that ice margin advance (or stillstand) was associated with a cold period, which was terminated upon incursion of warmer water into Davis Strait about 8300-8000 BP. Although not necessarily associated, it should be noted that this was also about the time of the breakup of the Laurentide ice sheet over Hudson Bay (Andrews and Falconer, 1969).

The middle Holocene records from the Baffin Bay region are difficult to characterize. The greatest concentration of driftwood in the Queen Elizabeth Islands, presumably implying the most open water, occurred between 6500 and 4200 BP (Fig. 26.9a; Stewart and England, 1983). The climatic optimum in the Clyde (Patricia Bay) pollen record at 6800-5700 BP is in partial agreement (Fig. 26.9d), but also coincides with ice margin advance, or at least reduced retreat rate, in central Baffin Island (Figs 26.9b, c, & e). The latter may perhaps be explained as a response of the ice sheet to the cessation of calving when it became wholly land-based (Andrews, 1973, 1982).

A readvance of the ice on Cumberland Peninsula at 5000-4500 BP (Fig. 26.9f), and reduced retreat rate in the Home Bay area at 5600-4100 BP (Fig. 26.9e), may be related to the severe cold period (Short et al., this volume) indicated by the Patricia Bay pollen record between 5700 and 4500 BP (Fig. 26.9d) though there is considerable uncertainty in the dating of the lower part of this core. By contrast the pollen evidence from Iglutalik Lake (Fig. 26.9g) is entirely contrary to the Patricia Lake record during this period and other evidence points to relatively mild conditions in Baffin Bay and southwestern Greenland at this time. For example, subarctic species of molluscs flourished on the east coast of Baffin Island between 5500 and 3500 BP (Fig. 26.9j; Andrews, 1972), and the climatic optimum in the pollen record from Frederikshab covers nearly the same interval, 5800-3200 BP (Fig. 26.9m). Other pollen records from southwestern Greenland indicate a generally warm climate from c. 8000 BP to at least 4000 BP. Eigenvector analysis of pollen-based temperature records from Baffin sland, Labrador and Keewatin also indicate relatively warm conditions from 5500 to 3500 BP (Andrews and Diaz, 1981).

THE LAST 6000 YEARS

Figures 26.10 and 26.11 show records covering the last 6000 and 1400 calendar years, respectively. For comparison with ice-core and tree-ring records, radiocarbon dates and the time scales of C-14-dated records have been converted to calendar years according to the dendrochronological calibration of Stuiver (1982), which covers the last 2000 years, and Ralph et al. (1973) for older dates. The calibration of Stuiver (1982) is probably very accurate, but another recent calibration by Klein et al. (1982) indicates that some of the earlier BC dates used here (i.e. according to the Ralph et al. 1973 curve) are about a century too old at times. However, the Klein et al. (1982) data are presented in a form which is not easy to use, and since no ice-core records before 300 BC are considered anyway, the older calibration is used. The radiocarbon date list and conversions are given in Appendix A.

Eight different kinds of paleoclimatic evidence are presented in Figures 26.10 and 26.11: (1) Wood (charcoal) north of present tree line, (2) transfer function reconstructions of summer temperature based on pollen assemblages, (3) lichenometric dates on moraines, (4) radiocarbon dates on various deposits which suggest climatic change, (5) proglacial sediment stratigraphy, (6) tree-ring width, (7) percent refrozen meltwater in ice cores, and (8) oxygen isotope ratios (0-18) in ice cores (cf. Table 26.1).

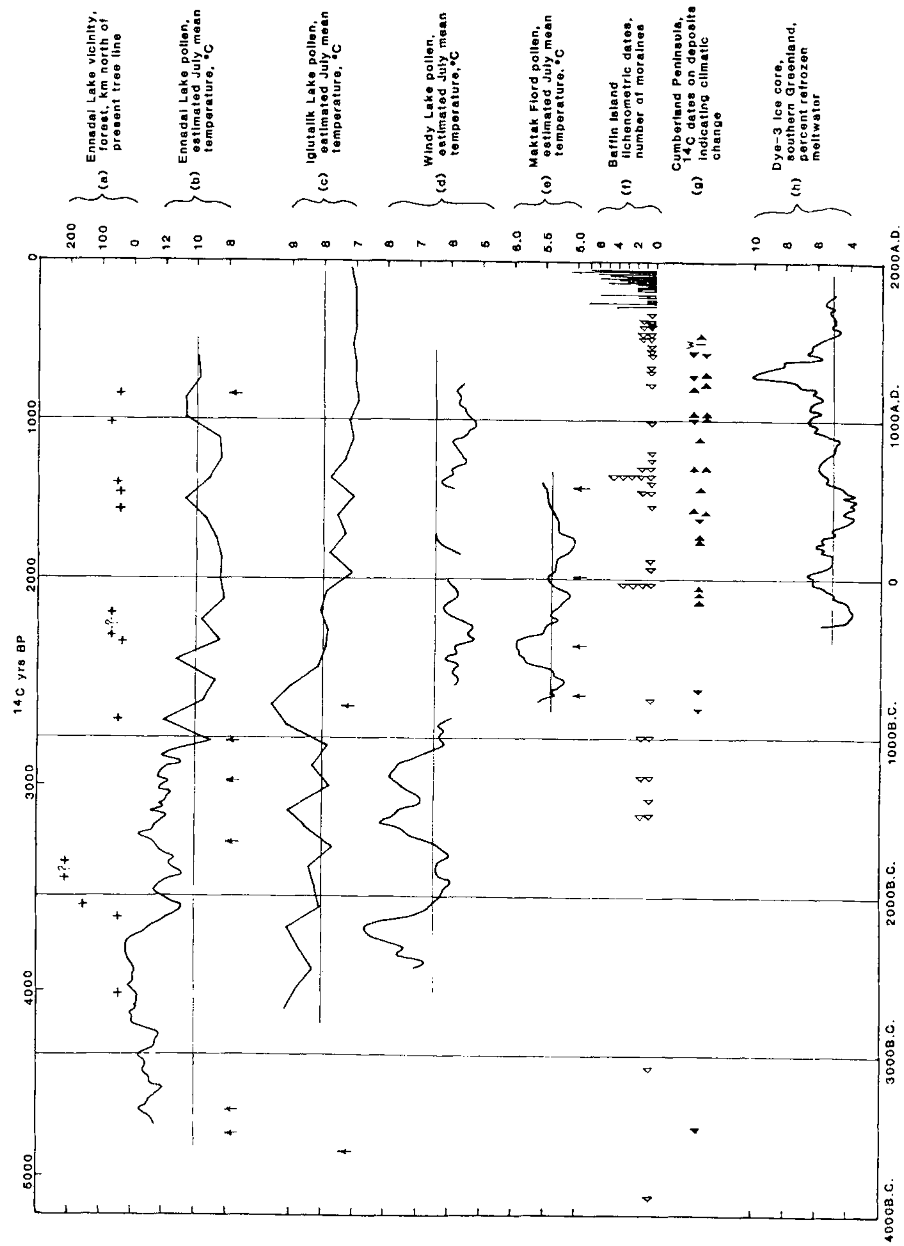
For the sake of comparability, and also for smoothing, the time series in Figure 26.10 have been reduced to approximate 120-year running means, 120 years being the approximate sampling interval of two of the records (Figs 26.10c and the upper part of 26.10b). At the top of Figure 26.10, two kinds of evidence from a location outside the Baffin Bay region, the vicinity of Ennadai Lake in Keewatin, are shown for comparison, both with each other and with the records from the Baffin Bay region. Back to about 1000 BC there is very good agreement between dates on fossil wood (charcoal) north of present tree line (Fig. 26.10a) and the pollen transfer function reconstruction of July temperature at Ennadai Lake (26.10b). However, the dates on the farthest northward extension of forest, at around 2000 BC, coincide with a relative low in the July temperature reconstruction, although it was still warmer than most of the post-1000 BC part of the record. The lifetime of this forest is unknown, but it may have existed through the long warm period before about 2200 BC indicated by the pollen record, and survived the subsequent cooling, until destroyed by fire. No wood from north of present tree line in this region has been reported which dates from later in the 2nd millennium BC (for which the transfer function also gives high temperatures), but there are several dates on spruce above present tree-line altitude in the Yukon from 3380-3050 C-14 years BP (c. 1900-1400 BC) (Rampton, 1971; Denton and Karlen, 1973).

The Keewatin records provide a well-dated and reasonably consistent climatic history with which to compare the data from the Baffin Bay region. The episode of relatively cool summers about 2100-1700 BC, which interrupted very warm intervals, is also found in the Baffin Island pollen transfer function results from Iglutalik Lake (Fig. 26.10c) and Windy Lake (Fig. 26.10d). However, the Baffin Island transfer function results show little resemblance after 1000 BP, other than long-term climatic deterioration. Their dating is uncertain to some degree, but it seems impossible to reconcile them entirely by any reasonable shift of their time scales. The part of the Iglutalik Lake profile (Fig. 26.10c) which has been analyzed by transfer function is not at all well-dated (Appendix A); the top of the section is assumed to be modern (Davis, 1980), but the long climatic quiescence after 900 AD which this implies, makes the assumption seem questionable. The Windy Lake record (Fig. 26.10d) has breaks where sand layers in the section were not sampled, which are interpreted by Andrews et al. (1980) as representing cold intervals. This section has many C-14 dates, but there is a fair amount of scatter in the depth-age data, and we have derived a time scale simply by a linear least-squares fit (see Appendix A).

The Maktak Fiord record (Fig. 26.10e) is well-dated, with four very consistent dates within 1300 years. However, much of the variation in the Maktak series (even in the non-smoothed data) is about the same as the standard error of estimate (about 1°C) of the transfer function used to derive it (Table 7 in Andrews et al., 1980). Indeed, this is true of much of the post-1000 BC variation in all three Baffin Island transfer function temperature reconstructions (note the difference in scales on Figs 26.10b, c, d, & e).

Some indication of climatic change on Baffin Island may perhaps be gained from lichenometric dates on moraines (Fig. 26.10f) and radiocarbon dates on various kinds of deposits which suggest transition from warmer to colder conditions or the converse (Fig. 26.10g). these data are drawn as triangles with the points toward the colder episodes (except for one date marked W which indicates warm conditions, and the lichenometric dates after 1700 AD, which are too numerous to portray in this way). The lichenometric dates from the vicinity of the Barnes Ice Cap and from Cumberland Peninsula are plotted together in Figure 26.10f, for they fall in similar clusters, using the same lichen growth curve for both (as suggested by Andrews and Barnett, 1979).

Figure 26.10 Data (and derived data) from the last 6000 calendar years. (a) Radiocarbon dates on fossil wood (charcoal) north of present tree line in Keewatin (Bryson et al., 1965; Bender et al., 1965, 1966, 1967). (b,c,d,e) Transfer function reconstructions of mean July temperature based on pollen assemblages in Keewatin and Baffin Island (Andrews et al., 1980; Andrews and Nichols, 1981; data supplied by J.T. Andrews). Reduced to approximate 120-year running means, except where the sampling interval is about 120 years. Arrows denote C-14 dates. (f) Lichenometric dates on Baffin Island moraines (Miller, 1973; Andrews and Barnett, 1979; Davis, 1980). Minimum dates for glacier advance. (g) Radiocarbon dates on deposits suggestive of climatic change, Cumberland Peninsula, Baffin Island (Miller, 1973; Dyke, 1977). Triangles point toward colder episodes. See Appendix B. (h) Percent refrozen meltwater in an ice core from southern Greenland (Herron et al., 1981), 120-year running mean.

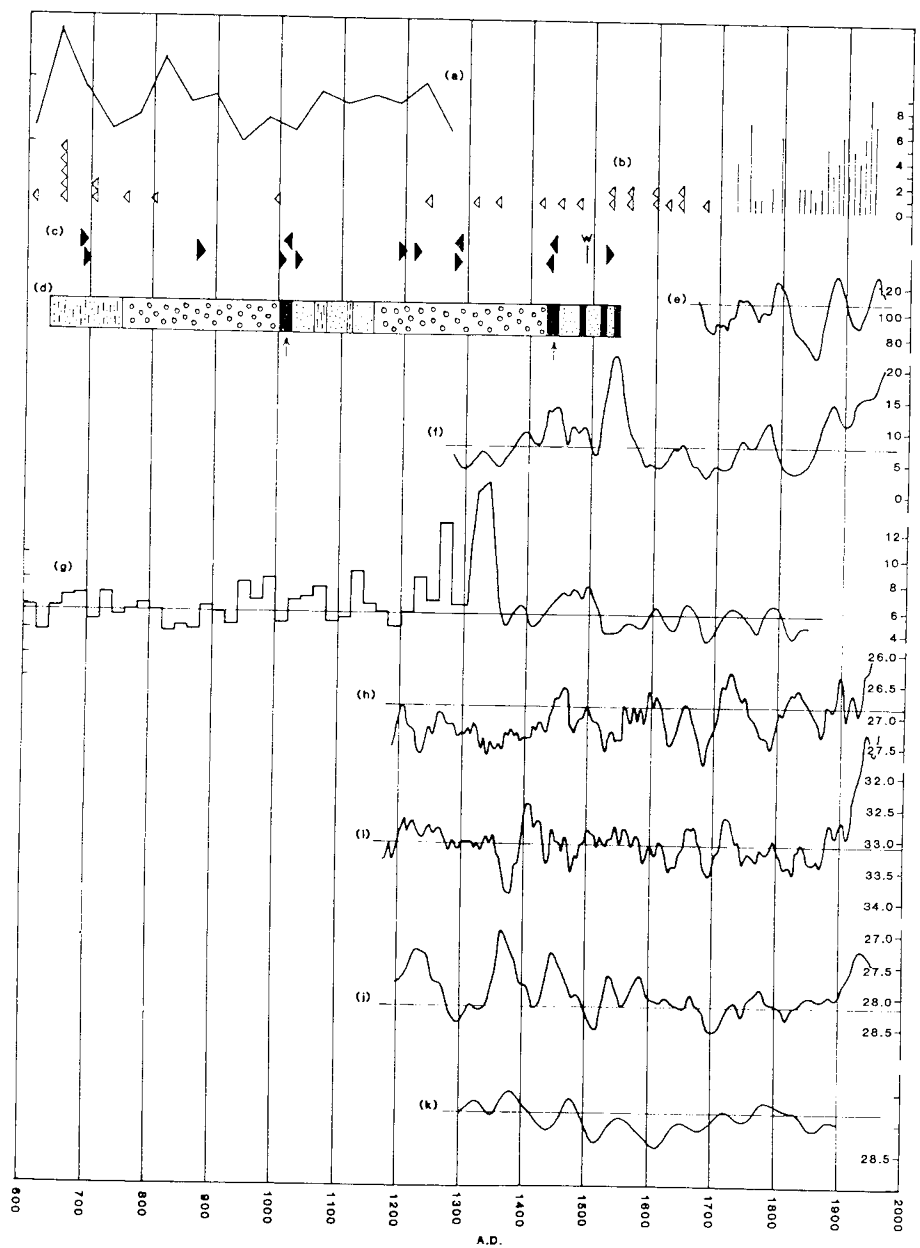


The oldest dates are widely scattered, and the lichenometric dates may be considerably in error, for the lichen growth curve is extrapolated from a control date at 500 BC. A cluster of lichenometric dates after 1500 BC indicates termination of a glacial advance which is probably related to the relatively cold period of 2100-1700 BC shown by the Windy Lake pollen data (the mean July temperature of about 6°C obtained for this interval by the transfer function is just below the 6.5°C it predicts from the modern pollen rain). Scattered lichenometric dates and C-14 dates on basal peat during the early 1st millennium BC are not clearly associated with the pollen transfer function results.

The cluster of lichenometric dates about 2000 years ago may be related to the cooling episode 400-100 BC in the Maktak (Fig. 26.10c) record (which amounts to 1.4°C in the non-smoothed data, and is therefore probably significant). The Dye-3 meltwater record (Fig. 26.10h) is also compatible with these lichenometric dates. The other Baffin Island data (Figs 26.10c, d, & g) are contradictory; however, C-14 dates (Fig. 26.10g) are on soils underlying eolian deposits (which Dyke, 1977, interpreted as a change to cold, dry conditions), and radiocarbon dates on soils can be centuries too old (Stuckenrath et al., 1979; Matthews, 1980). It is likely that the lichens grew on moraines deposited 200-100 BC, and the eolian sands were deposited during a later cold episode.

A large cluster of lichenometric dates after 500 AD (with a peak at 650 AD), together with two 5th century C-14 dates on glacial outwash (Fig. 26.10g; Appendix B), indicate a glacial advance sometime around the 5th century AD. This agrees very well with the low at that time in

Figure 26.11 The last 1400 years. (a) As Figure 26.10, but for individual samples at 40-year intervals. (b) As Figure 26.10f. (c) As Figure 26.10g. (d) Sediment stratigraphy in front of a Baffin Island glacier (Miller, 1973). Dashed - silt; fine stipple - fine sand; coarse stipple - coarse sand, gravel, cobbles; shaded - organic horizon. See text for interpretation. (e) Tree-ring width indices from vicinity of Fort Chimo, northern Quebec (Cropper and Fritts, 1981), 30-year running means. (f) Percent refrozen meltwater in two cores from the Devon Island ice cap (Fisher and Koerner, 1981), 30-year running means. (g) As Figure 26.10h, but 20-year means (up to AD 1300) and 30-year running means (after AD 1300). (h,i) 30-year running means of annual maxima of O-18 in Greenland ice cores, from (h) Milcent, and (i) Crete (data supplied by World Data Center A for Glaciology). (j,k) 30-year running means of mean annual O-18 in ice cores from (j) Devon Island (Paterson et al., 1977), and (k) Dye-3 Greenland (Herron et al., 1981).



the Dye-3 meltwater record (Fig. 26.10h), and with the interpretation that the contemporaneous sand layer in the Windy Lake section (Fig. 26.10d) represents a cold period. Variations in the Iglutalik Lake and Maktak Fiord records (Fig. 26.10c, e) during this period are not significant. The timing of this cold episode is worthy of some consideration, for if it occurred in the 5th century in the Baffin Bay region, then it was clearly opposite to climatic change in Keewatin (Figs 26.10a & b).

600 AD TO PRESENT

The records for the last 1400 years (Fig. 26.10) may be considered in somewhat more detail than the previous sets. The points on the Windy Lake transfer function results (Fig. 26.11a) represent individual samples at about 40-year intevals. As noted previously, this record is ambiguously dated, and this is especially so in the upper part of the section (Appendix A). However, the concentration of lichenometric dates (Fig. 26.11b) around the 7th century AD does imply a warming at that time, in agreement with the Windy Lake result (see previous section).

The set of C-14 dates on deposits indicative of climatic change (Fig. 26.11c and Appendix B), together with the sediment stratigraphy in front of a glacier (Fig. 26.11d), provide a consistent and fairly detailed history of climatic change on the Cumberland Peninsula of Baffin Island between the 7th and 16th centuries AD. The proglacial sediment stratigraphy is interpreted by assuming that gravel was deposited when the glacier was close to the site, and fine sediment or organic material when it was withdrawn. This differs from the original interpretaiton of Miller (1973), but agrees with most of the radiocarbon-dated deposits (Fig. 26.11c), and follows the interpretation of Patzelt (1974) for a similar proglacial sediment section in the Alps. It also agrees very well with the meltwater record from the Devon Island ice cores (Fig. 26.11f).

Later, in the 17th to 20th centuries, the Devon Island meltwater record is remarkably similar to the Fort Chimo (northern Quebec) tree-ring series (Fig. 26.11e). The Baffin Island lichenometric dates (Fig. 26.11b) are fairly consistent with the Devon Island record (Fig. 26.11f) after 1680. However, the scattered lichen dates in the preceding few centuries show no clear association with the other evidence from the eastern Canadian Arctic.

The 'Little Ice Age' minima in the Devon Island meltwater record at around 1600 AD, the late 17th century, and the early 19th century, are very similar in timing to those in many other proxy paleoclimate records from North America and Europe (Lamb 1977; Williams and Wigley, 1983). Prior to the late 16th century, however, there appear to be some important differences between climatic change in the eastern Canadian Arctic and the rest of North America and Europe.

Many proxy climate records from North America and Europe show a major transition from a cold episode in the 9th - 10th centuries (in places comparable in severity to the Little Ice Age) to a very warm episode in the 11th - 12th centuries or later (Williams and Wigley, 1983). In Europe, the latter is known as the 'Medieval Warm Period' (Lamb, 1977). The Windy Lake temperature reconstruction (Fig. 26.11a) and our interpretation of the proglacial sediment section (Fig. 26.11d) do imply warming from the 10th to the 11th century. However, the temperature increase at Windy Lake was less than the standard error of th transfer function used to derive it. Also, interpretation of the C-14-dated deposits (Fig. 26.11c and Appendix B) in the early 11th century is ambiguous, and only one moraine on Baffin Island has been lichenometrically dated to the 10th or 11th century. Thus, there may have been a slight climatic amelioration on Baffin Island at the time, but it appears to have been far less of a change than that implied by records from elsewhere in North America and Europe. Similarly, there was only a slight increase in snowmelt at Dye-3 (Greenland) at the time (Figs 26.10h, & 26.11g).

Summer climate in the eastern Canadian Arctic seems to have been even more anomalous in the 15th to early 16th centuries (Williams and Wigley, 1983). In the western United States, ring-widths of high-altitude, summer temperature-sensitive trees indicate a major cold period in the late 15th century, locally perhaps the most severe of the Little Ice Age (LaMarche, 1974; LaMarche and Stockton, 1974), and well-dated glacier advances occurred at the same time in the Yukon (Denton and Stuiver, 1966). Proxy climatic data from Europe also indicate a period of relatively cold summers in the 15th century (e.g. Schweingruber et al., 1978). In contrast, evidence from the esatern Canadian Arctic (Fig. 26.11c, d, & f) indicates that summers were relatively warm in the 15th and early 16th centuries, following a cold period in the 13th - 14th centuries. However, it must be noted that this evidence, although strong, does not receive support from lichenometric dates (Fig. 26.11b).

The date (1490) marked 'W' in Figure 26.11c is on organic material at the bottom of an ice wedge on Baffin Island, which implies seasonal thawing of ground perennially frozen at present. This was also the case in southern Greenland at the time, for material from burials of Norse settlers dating up to at least the late 15th century have been recovered from ground which was perennially frozen at the time of excavation in the early 20th century (Nörlund, 1924; Hovgaard, 1925). The good state of preservation of these materials suggested that the ground became perennially frozen not long after burial, and remained so until excavation.

These burials give general support to the meltwater record from the nearby Dye-3 ice core (Fig. 26.11g), which has high values during the time of Norse settlement from the late 10th century until the early 16th century, and generally low values thereafter. However, the Dye-3 meltwater record shows very little agreement with that from Devon Island (Fig. 26.11f); in fact the differences between the two in the 14th and 16th centuries are astounding. The two sites are a fair distance apart (Fig. 26.8) so it is possible that local climatic changes were quite different. Some difference is even perhaps to be expected, considering that melting on the Devon Island ice cap is strongly dependent on synoptic climate, especially on the frequency of Baffin Bay lows (Alt, 1978), which might have an opposite effect on melting at Dye-3. Yet we found that the Devon Island meltwater series bears a strong resemblance to the tree-ring series from distant Fort Chimo (Fig. 26.8) during the last few centuries, and the minima in the Dye-3 meltwater record (Fig. 26.10h) seem to be closely associated with clusters of lichenometric dates on Baffin Island moraines (Fig. 26.10f). The question of synchrony of climatic change even within the eastern Canadian Arctic - southern Greenland region remains unresolved, much less that over the entire Northern Hemisphere or globally (cf. Denton and Karlen, 1973).

An attempt has been made here to relate oxygen isotope data from two Greenland ice cores to summer temperature, by taking only the maximum O-18 values in an annual cycle. This was done for the Milcent (Fig. 26.11h) and Crete (Fig. 26.11i) cores, for the period of record since about 1200 AD when data on annual variations are available (date provided by P.K. MacKinnon, World Data Center A for Glaciology, 1981). As seen in Figures 26.11h & i, the exercise was not particularly informative, as the two records generally disagree with each other and with the Dye-3 meltwater record, except perhaps for some similarity between the Dye-3 and Milcent series (Figs 26.11g & h) from 1400 to 1700 AD.

Finally, the mean annual O-18 records from the Devon Island and Dye-3 ice cores (Figs 26.11j & k) are shown for comparison with the

summer melt records from the same locations (Figs 26.11f & g). In the case of Devon Island, there is a fair correspondence between the two curves after about 1650, and around 1300. However, between 1350 and 1600 the two curves show little resemblance. The mean annual O-18 record and melt record from Dye-3 (Figs 26.11g & k) are even less in agreement.

SUMMARY

There is little coherence in the pattern of early- to mid-Holocene climatic change in the Baffin Bay region when various records of ice-margin retreat, pollen, and mollusc assemblages are compared. The period of ice margin readvance on Baffin Island (and to some extent, in southwest Greenland) sometime between 8000 and 9000 C-14 years BP may have terminated synchronously with the opening of Davis Strait, the influx of subarctic species of molluscs on the east coast of Baffin Island, and the start of a long period of relatively warm-climate pollen assemblages in southwestern Greenland. Subsequently, a climatic optimum seems to have prevailed in most of the Baffin Bay region until about 3000 C-14 years BP, although interrupted in places by colder episodes.

For the period 4000 BP to present, pollen transfer function reconstructions of summer temperatures in Keewatin and Baffin Island indicate generally warm conditions up to the first millennium BC, and generally colder thereafter, although they differ greatly in detail. They do agree on a relatively cold interval within the earlier period, from about 2100-1700 BC, and this is supported by a cluster of lichenometric dates on Baffin Island moraines. However, a far-northward extension of treeline in Keewatin is also dated within the period 2100-1700 BC. After 100 BC, dates on wood north of the present treeline in Keewatin compare well with temperature peaks based on palynological (transfer function) reconstruction for the same vicinity at about 900 BC, 500-300 BC, 400-600 AD, and 1000-1200 AD. On Baffin Island, two episodes of glacier expansion, dated lichenometrically, compare very well with minima in a southern Greenland ice-core meltwater record at around 200 BC and 400-500 AD. The latter contrasts with the warm episode in Keewatin at that time.

Various kinds of evidence for climatic change in the eastern Canadian Arctic during the last 1400 years are reasonably consistent, but those from the Greenland ice cores disagree. The combined evidence from the eastern Canadian Arctic suggests that after a relatively warm 7th century AD, the climate deteriorated into the 10th century. Warmer

conditions then prevailed in the 11th and 12th centuries (as in Keewatin, and indeed elsewhere in North America and Europe), followed by a cooling to the 14th century. The 15th and early 16th centuries were warm; at times, at least as warm as at present. This contrasts with evidence (tree-ring and glacial) for a cold episode in western North America and Europe in the 15th century. The Devon Island ice-core meltwater record indicates a sharp cooling in the late 16th century. This record, the northern Quebec tree-ring record, and the Baffin Island lichenometric dates on moraines all show very good agreement from 1650 to the present. These indicate cold periods in the late 17th century and early 19th century, with warming during the 18th century and late 19th to 20th century (the latter interrupted by cooling around 1900).

It would be naive to suppose that the few cases discussed, in which two or more kinds of paleoclimatic evidence agree, provide a definitive history of climatic change in the region. Some may be fortuitous, and there are many other cases in which there is disagreement among different records. The inconsistencies may be due to misinterpretation, dating uncertainty, or differing responses to climatic change, as well as the possibility of local anomalies Paleoclimatic interpretation is only made possible by aggregation of different kinds of supporting evidence, and this survey shows that there is a need for much additional information to be collected in the Baffin Bay region.

APPENDIX A

SOURCES OF DATA, AND CONVERSION OF RADIOCARBON DATES TO CALENDAR YEARS

Figure No.	Reference
26.9a	Blake (1972)
26.9b, c, e, f, i	Andrews (1982)
26.9d, g, h	Short et al. (this volume)
26.9j	Andrews (1972), Laursen (1976), Miller (1980)
26.9k	Ten Brink and Weidick (1974)
26.9l, m, n, p	Fredskild (this volume)

Fig. No.	Calendar AD/BC	C-14 age BP	Lab. No.	Reference
26.10a	AD 1160	880 180	WIS-5	Bender et al. (1965)
		1050 180		
	AD 980	1140 90	WIS-17	"
	AD 600	1450 90	WIS-15	"
	AD 540	1530 80	WIS-96	Bender et al. (1966)
	AD 430	1590 80	WIS-37	Bender et al. (1965)
	220-360 BC	2140 80	WIS-136	Bender et al. (1967)
	400 BC	2210 160	WIS-29	Bender et al. (1965)
	890 BC	2670 105	WIS-93	Bender et al. (1966)
	1790-1890 BC	3430 110	WIS-12	Bender et al. (1965)
	2060 BC	3540 110	WIS-52	Bender et al. (1966)
	2140 BC	3650 100	WIS-80	Bender et al. (1966)
	2620 BC	4000 160	WIS-7	Bender et al. (1965)
26.10b	AD 1160	870 60	Gak-5062	Nichols (1975)
	1020 BC	2790 100	Gak-5061	"
	1270 BC	2960 100	Gak-5060	"
	1660 BC	3340 120	Gak-5059	"
	3350 BC	4520 110	Gak-5057	"
	3500 BC	4690 140	GSC-1781	"
26.10c	795 BC	2565 190	GX-6292	Davis (1980)
	3610 BC	4765 200	GX-5625	"
26.10d	AD 1300	640 155	Gak-5449	Davis (1980)(mean depth 2.5 cm)
	AD 1030	960 200	Gak-5450	" (7.5 cm)
	AD 1200	850 65	DIC-327	" (15.5 cm)
	AD 560	1500 85	DIC-390	" (62 cm)
	AD 10	1990 180	Gak-5411	" (69 cm)
	130 BC	2060 85	Gak-5412	" (90 cm)
	730 BC	2470 390	DIC-515	" (99 cm)
	1060 BC	2825 65	SI-2950	" (128 cm)
	1410 BC	3070 75	DIC-402	" (157.5 cm)
	2140 BC	3650 200	SI-2556	" (205 cm)
	2050 BC	3525 60	SI-2951	" (217.5 cm)
Least-squares fit: Date (AD) 1273 = 16.545 x depth(r^2 = .97)				
26.10e	AD 580	1480 160	Birm-370	Boulton et al. (1976)
	AD 25	1970 200	Birm-535	"
	410 BC	2240 190	Birm-536	"
	765 BC	2500 170	Birm-380	"
26.10f	Historical dates			Miller and Andrews (1972)
	AD 1290	680 80	Gak-3722	Miller (1973)
	500 BC	2400 90	Gak-1992	Miller and Andrews (1972)
26.10g	See Appendix B			
26.11a	See 26.10d			
26.1b	See 26.10f			
26.11d	AD 1435	450 130	Gak-3726	Miller (1973)
	AD 1010	1010 100	Gak-3725	"

Radiocarbon dates on deposits which indicate climatic change, Cumberland Peninsula, Baffin Island

Lab. No.	C-14 yrs BP	AD/BC	Description of deposit	Interpretation
Gak-3099	330±90	AD 1520	Dead moss in area of lichen kill	Cold after
Gak-2983	350±100	1490	Organic debris in bottom of ice wedge	Warm
Gak-3357	430±90	1440	Soil overlain by peat	Warm after?
Gak-3726	450±130	1435	Organic horizon underlain by outwash	Cold before
Gak-3098	680±90	1290	Base of peat section	Warming?
Gak-2792	730±70	1280	Top of peat section	Cooling?
Qu-305	830±70	1215	Soil overlain by eolian sand	Cold after
Gak-3094	850±110	1190	Top of peat overlain by eolian sand	Cold after
BGS-267	970±80	1025	Soil overlain by eolian sand	Cold after
Gak-4839	970±70	1025	Organic loess overlain by eolian sand	Cold after
Gak-3725	1010±100	1020	Soil underlain by outwash	Cold before
SI-2550	1025±100	1000	Dead moss in area of lichen kill	Cold after
Qu-301	1170±150	870	Peaty soil overlain by outwash	Cold after
Gak-3160	1260±150	690	Soil buried between moraine crests	Cold after
Gak-4307	1290±100	685	Soil overlain by loess	Cold after
BGS-268	1500±80	560	Soil overlain by eolian sand	Cold after
Qu-307	1610±230	425	Peaty soil overlain by outwash	Cold after
Qu-303	1640±130	410	Peaty soil overlain by outwash	Cold after
Gak-2575	1670±90	390	Base of peat section	Warming?
SI-1703	1740±70	260	Soil overlain by eolian sand	Cold after
GSC-2084	1790±80	230	Soil overlain by eolian sand	Cold after
SI-1700	2015±60	60 BC	Soil overlain by eolian sand	Cold after
SI-1702A	2025±105	95 BC	Soil overlain by eolian sand	Cold after
Gx-3271	2080±190	150 BC	Soil overlain by eolian sand	Cold after
BGS-269	2450±90	660 BC	Base of peat	Warming?
SI-2555	2570±75	800 BC	Base of peat	Warming?
SI-1699	4660±90	3440 BC	Base of peat	Warming?

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Survey – Part V

The chapters in the final section of this volume deal largely with th history of events during the last 10,000 years. This period includes the retreat of ice sheets and glaciers from the eastern Canadian Arct and West Greenalnd during early and middle Holocene time, but it also includes the readvances of local glaciers during the geologic-climate interval called the neoglacial. However, it should be noted that chapters in other sections of this volume deal with glacial, climatic and oceanographic events during the Holocene. Thus chapters 15, 16, 18, and 19 contribute substantial information to the broad topic of Holocene glacial and climatic events in the area surrounding Baffin Bay.

The chapter by Quinlan (Chapter 20) represents a significant contribution to our attempts to portray the changing character of the late Quaternary North American ice sheet. The maps and figures provided by Quinlan represent an important series of hypotheses agains which field observations can be tested. This in turn will allow a re-evaluation of the glacial isostatic model that Quinlan and others have employed. A preliminary test of the model(s) in Frobisher Bay indicates some areas of disagreement. However, the Frobisher Bay area has largely been worked on during large-scale reconnaissance surveys where the emphasis has been on mapping of glacial and glacial marine sediments (see Chapter 18).

Chapters 22, 23, 24, 25, and 26 deal with various aspects of Holocene climate. The detailed investigation of neoglacial moraines by Davis (Chapter 24) indicates that the fluctuation of glacier snouts in southern Cumberland Peninsula has a much higher frequency content than the proposed 2500 year cycle. However, the pollen records from both the eastern Canadian Arctic and West Greenland (Chapters 22 and 23) illustrate broader amplitude oscillations of climate.

A comparison of these two chapters indicates a substantial difference in the degree to which palynology has been used as a tool for reconstructing Holocene vegetation and climate histories. In Greenland there has been a long tradition of palynological studies and the reader must be impressed by the wealth of detail in Fredskild's chapter. This detail extends to a recognition of local pollen taxa that is virtually unparalleled in North American studies. In contrast, the beginning of palynological studies in the Eastern Canadian Arctic have tended to focus more on the climatic implication of the pollen record and less on the history of vegetation succession. However, a theme that might link and serve ...