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Climatic Evolution of the Eastern Canadian Arctic and Baffin Bay During the Past Three Million Years [and Discussion]

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Climatic evolution of the eastern Canadian Arctic and Baffin Bay during the past three million years

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[Plates 1 and 2]

The outer east coast of Baffin Island is characterized by a series of sedimentary forelands. These contain a variety of litho- and biofacies associated with glacial marine and marine deposition into sea levels higher than those of the present. These high relative sea levels were associated with glacial isostatic loading and unloading of the crust by the NE sector of the Laurentide Ice Sheet. On the basis of amino acid epimerization ratios, eleven chronologically distinct units are delimited. The youngest unit is less than or equal to 10 ka, but all others are at or beyond the limits of radiocarbon dating. Based on biostratigraphy and amino acid data, the oldest units exposed in the forelands may be Pliocene in age. Molluscs, Foraminifera, and the palynology of buried soils and organics, indicate that the vast bulk of the exposed sequences contain floras and faunas that represent environments *warmer* than those at present. An analysis of modern and fossil pollen spectra suggests a steady decrease in low arctic conditions throughout the Quaternary.

Introduction

The Neogene palaeoclimatology, palaeoceanography, and glacial geology of Baffin Bay, the northern Labrador Sea, and the adjacent land masses (figure 1) is undoubtedly an important, if only partly understood, element in the Pliocene and Quaternary history of the Northern Hemisphere. Keen (1980) demonstrated that in the Northern Hemisphere the mean summer (June–August) temperature at 70° N is closely correlated with the summer temperature of the area between West Greenland and eastern Baffin Island (period A.D. 1951–76). This implies that the zonal average temperature is being primarily forced by the changes that occur over this small ocean basin, which links the Arctic Ocean to the main western North Atlantic (figure 1). At the hemisphere scale (Barry *et al.* 1975), the transect between West Greenland and the eastern Canadian Arctic coastline represents one of the most extreme July temperature gradients in the Arctic, with a difference in temperature of 4–6 °C across the bay (Williams & Bradley 1985).

In several publications attention has been drawn to the contrasts in climate, ice extent and nearshore oceanography between West Greenland and eastern Baffin Island (see, for example, Jacobs *et al.* 1985; Andrews *et al.* 1981) and this has been used to challenge the conventional mid-latitude orthodoxy that the cause of glaciation is a matter of a depression in temperature, with no explicit account being made of the role of snow accumulation.

The role of oceanography in the present and past climatic changes in the area is probably profound. Under present conditions (figure 2) the oceanography is dominated by a

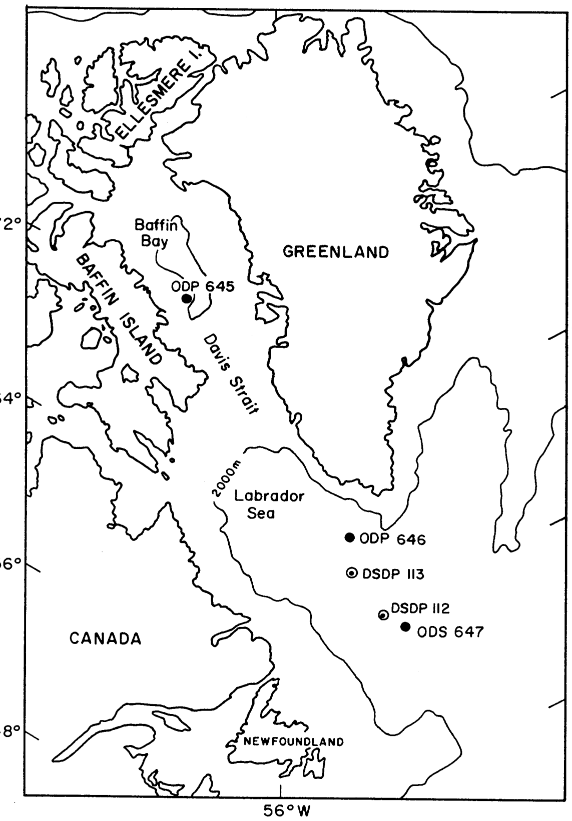

FIGURE 1. The 2000 m contour in the NE North Atlantic and the location of DSDP and ODP cores. The latter were taken on leg 105 in the autumn of 1985.

counterclockwise gyre: 'warm' water moves northward along the coast of West Greenland as the West Greenland Current (WGC). The main current turns westward in the north of Baffin Bay and descends to a depth of about 200 m, where it is held offshore against the narrow shelf of eastern Baffin Island. A shallow but cold current, the Canadian Current (CC), flows south along the west coast of Baffin Bay. The CC originates from the Arctic Ocean and moves southward through the arctic channels of the High Canadian Arctic. In the vicinity of Hudson Strait the CC joins with water from Hudson Bay and Foxe Basin and becomes the Labrador Current (LC). Both the CC and the LC are important mechanisms for the southward drift of icebergs from West and Northwest Greenland along the Canadian coastline.

Figure 2 shows the boundary between the marine arctic–subarctic and terrestrial high-arctic–low-arctic environments. This is controlled by the interactions of the WGC and CC. Diagnostic species of the low-arctic and subarctic zones are dwarf birch (*Betula* spp.) and the molluscs *Mytilus edulis* and *Macoma balthica*, among others (see Andrews 1972; Andrews *et al.* 1981; Lubinsky 1972, 1980). The average July temperature at the arctic–subarctic boundary is *ca.* 6 °C.

The west–east oceanographic contrast of Baffin Bay is also illustrated by a major contrast in the extent of glaciation between the Greenland and Canadian sides of the bay. The Greenland Ice Sheet has probably been a stable element in the geography of the region for most of the Quaternary. On the other hand, the northeastern sector of the Laurentide Ice Sheet reached tidewater along the Labrador and Baffin Island coasts several times during the same period,

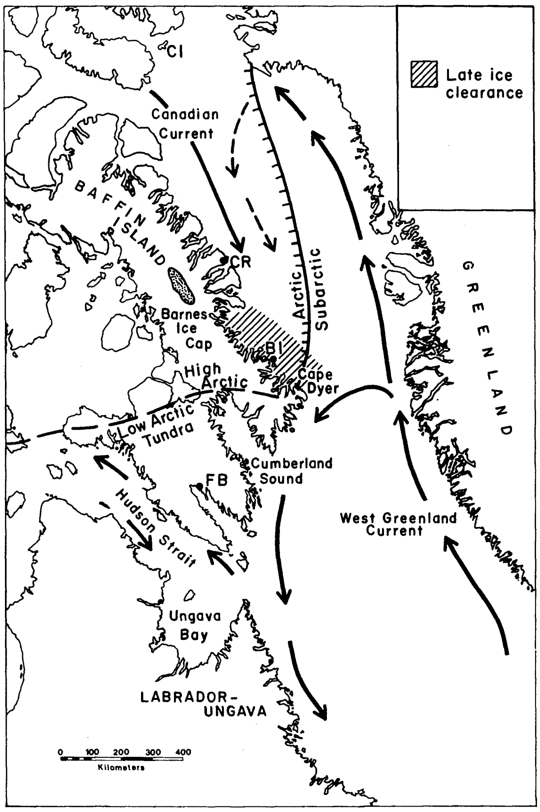

FIGURE 2. Generalized oceanographic circulation in Baffin Bay and the Labrador Sea. The arctic–subarctic zoogeographic boundary is shown (Dunbar 1968); this also corresponds with the northern limits of dwarf birch species. CR, Clyde River; BI, Broughton Island; FB, Frobisher Bay.

but today there are only a few small (less than 6000 km^2) relics of this once massive ice sheet. Despite the relatively small area of ice on Baffin Island today (37000 km^2), the remark by Tarr (1897) in the late 19th century, that Baffin Island is indeed 'wonderfully close to glaciation', is still appropriate.

Objectives

Of the many topics that might be evaluated, I wish to focus on an assessment of the chronology of glaciation and the palaeoenvironmental conditions associated with glaciation. Information from Ocean Drilling Project (ODP) leg 105, sites 645, 646, and 647 (figure 1), which were drilled in 1985, will add considerably to our knowledge of Quaternary events. The results of these surveys are just now being presented at meetings, but I would caution that evidence from these sites must be critically evaluated against the existing terrestrial evidence, much of which will be reviewed in this paper.

GLACIATION AND PALAEOENVIRONMENTS

The evidence from DSDP sites 112 and 113, between Greenland and Newfoundland (figure 1), were initially used to suggest that the onset of glaciation occurred *ca.* Ma BP (Berggren 1972). Relatively warm faunas occurred in the Labrador Sea during parts of the Pliocene (Thunnel & Belyen 1981) but in the earliest late Pliocene the Labrador Sea was invaded by polar waters (Poore & Berggren 1974). Evidence from Baffin Bay (Shipboard Party 1986) suggests that cooling may have started by 8 Ma BP. The pollen evidence from ODP site 645, in Baffin Bay, indicates that the transition from boreal forest to tundra occurred during the upper Pliocene (de Vernal *et al.* 1986). Srivastava *et al.* 1986) noted that the onset of glaciation may have begun as early as 3.4 Ma BP in Baffin Bay and 2.5 Ma BP in the western North Atlantic. These estimates are in agreement with data from the northeast Atlantic, where the surface water changed from warm-temperature to subpolar between 3.4 and 2.0 Ma BP (Loubere & Moss 1986). I once proposed (Andrews 1974) that glaciation of Greenland commenced in the Miocene, but no proof for this proposal is available.

The terrestrial evidence for late Pliocene climates is fragmentary on both Baffin Island and Greenland. In East Greenland, the Lodin Elv Formation is placed at the Pliocene–Pleistocene transition (Feyling-Hanssen *et al.* 1983). In North Greenland, the Early Quaternary Kap Kobenhavn Formation (Funder *et al.* 1984) records iceberg deposition *ca.* 1.8 Ma BP. These latter sediments contain an arctic molluscan fauna which is overlain by sediments with a boreal–low-arctic fauna and flora with affinities to communities which currently lie over 2000 km to the south of the site.

On the west side of Baffin Bay, the first discovery of terrestrial interglacial sediments was made by Andrews in 1962 along the Isortoq River, north central Baffin Island; in 1963 additional interglacial sediments were noted around the northern flank of the Barnes Ice Cap (figure 3). Terasmae *et al.* (1966) reported that the plant macrofossils and pollen assemblages indicated a low-arctic environment with the nearest present plant analogues some 600 km south of the Isortoq site. Radiocarbon dates gave variable ages in the range 14 to over 40 ka, but the younger dates were ascribed to the reworking of the sediments during the late Holocene retreat of the proto-Barnes Ice Cap. These plant-bearing beds were ascribed to the Flitaway Interglaciation, which was considered to be last-interglacial in age.

The extensive forelands that extend as low plains from the mountains toward Baffin Bay (figures 4 and 5) are a critical depository of Quaternary geological information. Only in western Spitsbergen are there comparable extensive Quaternary sections (see, for example, Boulton *et al.* 1982) along formerly glaciated coastal margins. Goldthwaite (1950), Loken (1966) and Feyling-Hanssen (1976*a*) were the first researchers to study these wave-cut exposures, which extend for more than 300 km along the outer east coast of Baffin Island (figure 4). Where cliffed by wave erosion (figure 5), the exposures consist of sediments associated with successive glacial isostatic sea-level oscillations (Feyling-Hanssen 1976*a*, *b*, 1985; Miller *et al.* 1977; Nelson 1981; Brigham 1983; Mode *et al.* 1983). The exposed sediments consisted primarily of (less than 100 m depth of water) marine silts and clays, littoral sands and gravels, some till, and peat and soils. The first ^{14}C dates (Loken 1966) indicated that the great bulk of these sediments were older than 50 ka. Research on these sections has focused on (i) the nature of the sediment cycles, (ii) their ages, and (iii) their faunas and floras.

(*a*) *Sediment cycles*

Miller *et al.* (1977), Nelson (1981) and Brigham (1983) described the succession of sediments exposed in the cliff sections (e.g. those in figures 4 and 5). From these data, Mode *et al.* (1983) presented a generalized facies model and linked the sediment sequence with the combined interactions of changes in glacier extent and glacial isostatic changes in relative sea level. Figure 6 is a schematic sketch of an idealized sequence for one glacial–interglacial cycle. The

FIGURE 4. Oblique air photograph looking northward along the outer coast of eastern Baffin Island. In the foreground is the Qivitu foreland. Numbers 1, 4 and 5 point to specific wave-cut sections: 2 locates a small raised delta some distance below the local marine limit; and 3 points to the right lateral moraine of a former ice lobe moving seaward down Narpaing Fiord.

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FIGURE 5. Photograph of the exposed section at number 5 (figure 4), showing the freshly eroded face of the wave-eroded section.

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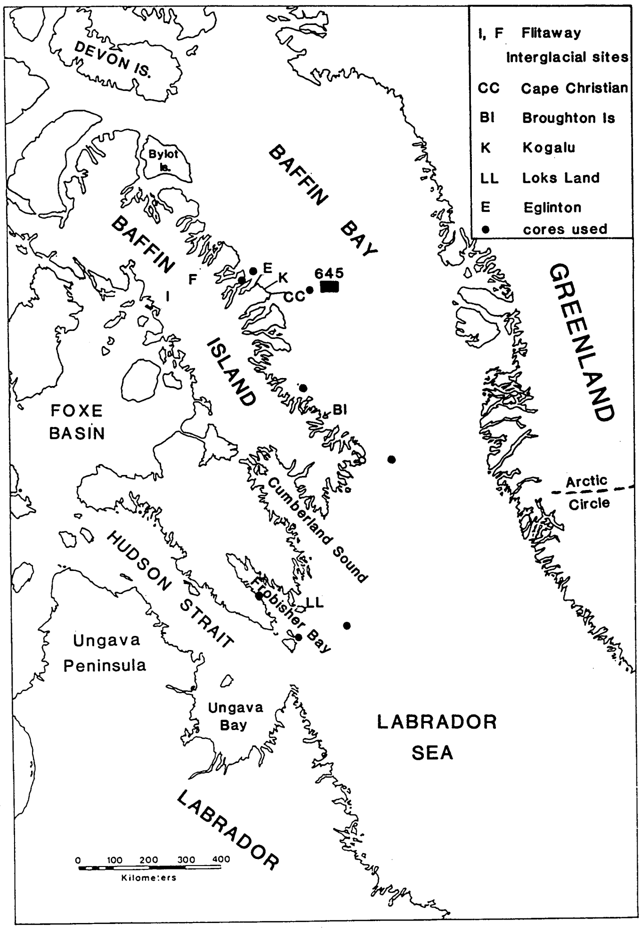

FIGURE 3. Location of typesites for some of the units described in the paper. Capital letters are used to locate the sites: I, Isortoq; F, Flitaway; CC, Cape Christian; K, Kogalu; BI, Cape Broughton; LL, Loks Land; E, Eglinton. Piston core coverage shown.

diamictons represent either basal tills or pebbly glacial marine sediments. In the eastern Baffin Island sequences, the diamictons frequently contain well-preserved foraminifera and whole bivalves, and are glacial marine sediments deposited close to ice margins that reached the outer coast and extended onto the continental shelf.

The vast majority of the forelands consist of a variety of shallow-water and littoral sediments deposited during marine transgressions and regressions. Faunas from these sediments, mainly molluscs and Foraminifera, can be used to deduce palaeoenvironmental conditions during each glacial–interglacial cycle. In some instances, organic sediments are preserved (O facies (Mode *et al.* 1983; Mode 1980, 1985)) and record periods when the sea level was at or *below* the land surface (figure 6). Such intervals may represent periods of global non-glacial climate, but periods of low sea level (and non-deposition) might also occur at times when regional glacial extent was reduced and global sea level was low. Such conditions would apply if there were an out-of-phase relationship between glaciation at high and lower latitudes (cf. Andrews & Miller 1984; Boulton *et al.* 1985; Morner 1977).

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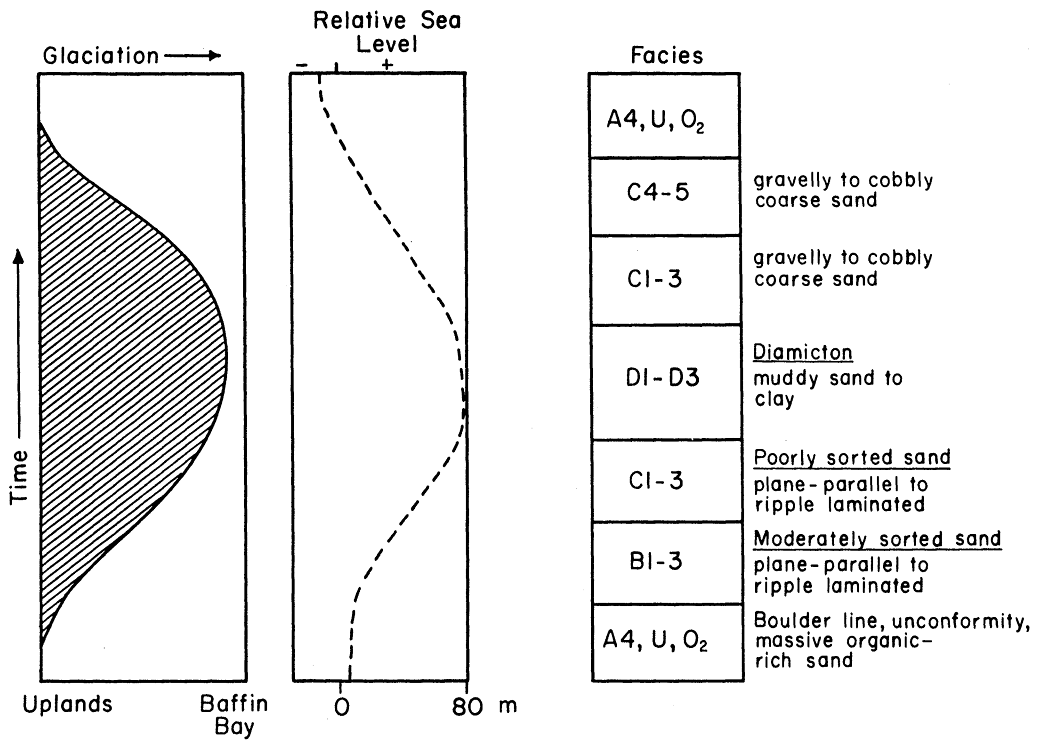

FIGURE 6. Sketch of the association between glaciation, relative sea level rise, and facies exposed in the forelands (e.g. figure 4) (after Mode *et al.* 1983) (see also figure 9).

(*b*) *Age of the sediments*

Because the bulk of the sediments lie beyond the limits of ^{14}C dating (Szabo *et al.* 1981) a variety of other dating methods have been used in an attempt to define the age, and number, of sediment packages. U-series analyses of marine molluscs proved useful in providing a series of *minimum* ages for several units. However, amino acid epimerization of common arctic bivalves, such as *Mya truncata* and *Hiatella arctica*, has been used successfully to delimit a series of aminozones along the length of the east coast of Baffin Island (see, for example, Miller *et al.* 1977; Nelson 1982; Brigham 1983; Mode 1985). This extensive body of data was reviewed by Miller (1985).

Amino acid ratios for D-alloisoleucine to L-isoleucine (aIle: Ile) in both the free (F) and total (T) fraction, increase to an equilibrium value of 1.3. The rate of epimerization is a function of both the thermal history of the site and the time since death of the organism. The aIle: Ile ratio can be used as a correlation tool for disjunct stratigraphic units. With some independent geochronological age control, the amino acid ratios can be used to estimate a range of probable ages, based on approximations of 'climate' (i.e. ground temperature). McCoy (1987) discussed errors associated with these various applications (see also Miller & Mangerud 1985).

Miller (1985, Table 14.11) evaluated the amino acid data for the late and middle Quaternary, whereas Mode (1985) presented evidence on older units (figure 7). A consideration of various thermal models suggested that the most likely effective diagenetic temperature (EDA) was -9 °C, which is associated with a mean annual temperature (MAT) of -11 to -15 °C (the present MAT is *ca.* -12 °C). The chronology shown in figure 7 is based on this model. Of note is the large number of distinct aminozones. Eleven aminozones are

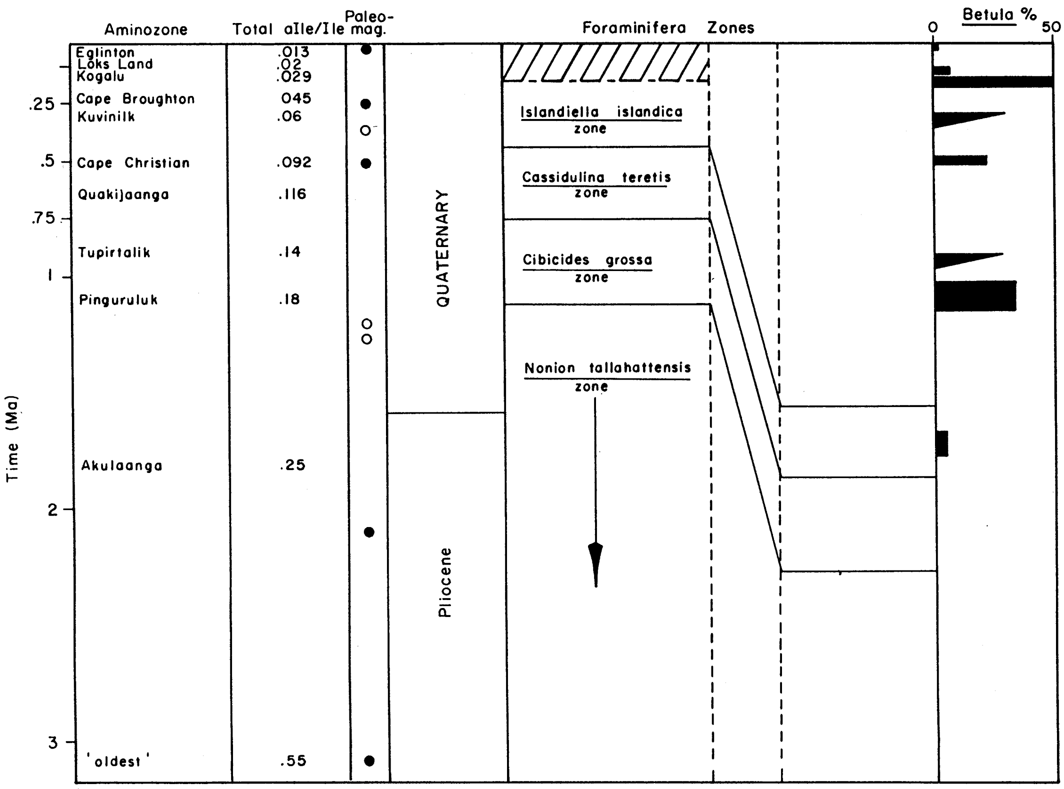

FIGURE 7. Possible geochronology for the Baffin Island aminozones (after Miller 1985) showing the limited palaeomagnetic results (solid circles represent normal inclinations). Variations in the percentages of *Betula* pollen (right-hand column, from Mode (1985)) represent a terrestrial subarctic element. Ascription of the foraminiferal zones depends on either the amino acid data (left, Miller 1985; Mode 1985) or on the biostratigraphy (Feyling-Hanssen 1985).

delimited, of which only the Eglinton and Loks Land are less than 40 ka old; this suggests that the preservation potential for these 'glacial isostatic facies' (Miller *et al.* 1977; Boulton *et al.* 1982) is remarkably high. This record in itself suggests that glacial ice was not an effective agent of sediment removal along the forelands of outer Baffin Island, although it must be stressed that the aminozones are *not* preserved anywhere in their entirety (cf. Gibbons *et al.* 1984).

In an initial chronology for eastern Baffin Island, Miller *et al.* (1977) used a U-series date to correlate the Cape Christian aminozone with the peak of the last interglaciation. Thus the Kuvinilk and Kogalu aminozones were placed in marine isotope stage 5. The Cape Christian and Flitaway deposits were correlated on the basis of their pollen assemblages. With a better understanding of both the amino acid epimerization rates, and further U-series determinations (Szabo *et al.* 1981; Brigham 1983), it was clear that the Cape Christian aminozone pre-dated stage 5e (i.e. it was much older than 125 ka).

The oldest aminozone exposed at the base of the forelands has a mean aIle: Ile ratio (T) of 0.55 (Miller 1985, p. 417). Based on the preferred EDA, this indicates an age of 3.5 Ma for this unit; but it could be as young as 1.6 Ma if an EDA of −5 °C is adopted. Limited palaeomagnetic data (Jacobs *et al.* 1985, figure 2.4) gave normal inclinations for the oldest sediment (figure 7). This result does not contradict an age of 3 Ma for the oldest aminozone, but the age estimates on figure 7 have error bars of ±50%!

Feyling-Hanssen (1985), Mode (1985), and Andrews *et al.* (1981) have reviewed the faunal and floral data preserved in the coastal foreland sediments. Marine molluscs, Foraminifera, and pollen provide information on the palaeoclimatology during different intervals of deposition of the 'glacial isostatic facies'. The arctic–subarctic faunal and high-arctic–low-arctic floral boundary (figure 2) represents an 'on–off' ecostratigraphic switch that is sensitive to the relative strengths of the WGC or CC along the coast of Baffin Island. However, the Foraminifera also have biostratigraphic significance, which is elaborated by Feyling-Hanssen (1985, pp. 366–369, Figure 13.18). On the basis of biostratigraphy he assigns the three lowest foraminiferal zones (figure 7) to the Pliocene, and correlates them with the Lodin Elv Formation of East Greenland. Mode (1985, Table 17.1) compared his aminozones from the Clyde Foreland with Feyling-Hanssen's (1985) foraminiferan biostratigraphy (note: Mode (1980) sampled the same units for amino acid purposes). If Feyling-Hanssen's biostratigraphy is used to position the Pleistocene–Pliocene boundary, this results in a 'long' chronology, as the Cape Christian aminozone dates from *ca.* 0.75 Ma BP and would require a long-term EDA of −11 °C (Miller 1985, Table 14.11; Mode 1985, Table 17.1) (figure 7). Such a result cannot be excluded with present information.

(*c*) *Palaeoenvironments during deposition*

The palynology of Baffin Island interglacial deposits (Terasmae *et al.* 1966; Miller *et al.* 1977; Mode 1985; Short *et al.* 1985) indicates that, most frequently, they consist of Gramineae–*Salix–Betula*) assemblages. Characteristically, the interglacial samples from north and east-central Baffin Island contain more *Betula* pollen than Holocene samples from the same area. On this basis, the present interglaciation is more severe than several earlier interglacial periods (see Mode 1985, Figure 17.3) (figure 7). Figure 8 is a comparison of Holocene pollen assemblages from northern Baffin Island with those from some of the older units (see figure 7). The diagram is based on a cluster analysis of the modern surface pollen rain (see, for example, Short *et al.* 1985) and the pollen spectra of peats and soils from various O facies (figure 7) within the cliffs. The north–south widths of the units (figure 8) represent the range of pollen assemblages along this 1500 km transect, whereas the width is a measure of the probability of the fossil sites being similar to the modern pollen spectra along the transect. This diagram shows that, even during the Holocene local thermal maximum, conditions were not as equable as they were during some earlier non-glacial intervals.

Of key concern is the stratigraphic position of the Flitaway and Isortoq plant-bearing beds (figures 3 and 8). These beds contain macrofossils of *Betula* and *Ledum groenlandicum*, and sufficient *Alnus* pollen to indicate that alder (probably *Alnus crispa*) grew close to the site (Terasmae *et al.* 1966). Collections from these units have been sampled for plant fossils (M. Kuc, personal communication, 1983); deposited in the National Type Collection of the Geological Survey of Canada) and beetles (Morgan *et al.* 1988). Both the moss flora and the insect fauna suggested that the sites were located close to, or even just within, the treeline. Morgan *et al.* (1982) estimated that the insect fauna indicated a July temperature 4–6 °C *above* that of the present, that is close to 10 °C. This would place the estimated July temperature 1–4 °C *above* the estimated Holocene thermal maximum for Baffin Island (Short *et al.* 1985).

When during the Neogene were temperatures this high over the Eastern Canadian Arctic? Some evidence occurs in the palynology of long piston cores from Baffin Bay and the evidence from the ODP site 645 (figures 1 and 3). de Vernal *et al.* (1986) noted that site 645 the transition

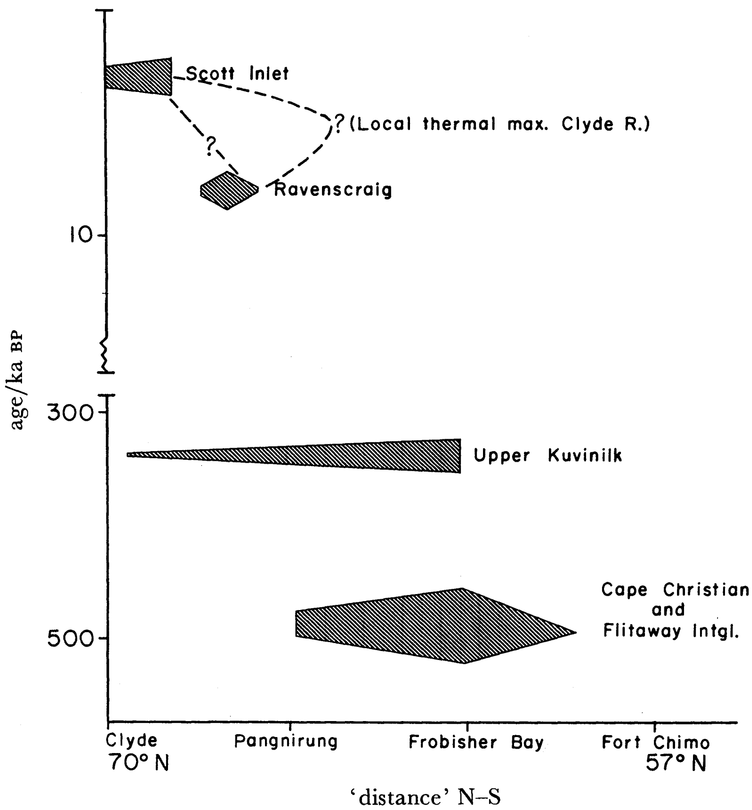

FIGURE 8. Position of fossil pollen spectra from sites on Baffin Island (figure 3) in the cliff sections near Clyde River and from the Flitaway sediments, compared with modern surface pollen rain on a transect from the boreal–tundra ecotone northwards to the Clyde River (figure 3).

from boreal–low-arctic vegetation to tundra occurred in the late Pliocene. Mudie & Short (1985, Figure 10.10) showed that, at a site south of site 645, subarctic shrub (*Betula*, *Alnus*) and conifer (*Picea* and *Pinus*) pollen occurred throughout the last 0.3 Ma and were advected to the area by either wind or ocean currents. Their graph of tree pollen percentages declines, with fluctuations, from a peak during marine-isotope stage 7 to low values within stage 1. Because pollen grains in deep-sea sediments are far-travelled, it is difficult to relate these assemblages directly to those on adjacent land masses; nevertheless, the boreal forest–tundra ecotone character of the Flitaway interglacial beds suggests considerable antiquity and certainly an age of much more than 125 ka. A late-Pliocene–early-Quaternary age cannot be ruled out, and they may be correlated with the Kap Kobenhavn floras of North Greenland (Funder *et al.* 1984).

The molluscan faunas in the raised marine sediments of eastern Baffin Island have been described qualitatively by Andrews (1972), Andrews *et al.* (1981) and Mode (1985). The northern limit of subarctic species, such as *Mytilus edulis*, occurs about 800 km farther north, in West Greenland, than it does on eastern Baffin Island (figure 2). In addition to species that extend to the limit of the subarctic zoogeographic province there are at least two other species that are currently restricted to the area of south of Hudson Strait; these are the gastropod *Colus spitsbergensis* and the pelecypod *Venericardia borealis*. The latter has only been found (one or two valves only) in some of the basal foreland sediments associated with total aIle : Ile ratios of 0.22.

This places the unit in the *Nonion tallahattensis* zone (figure 7) which may be late Pliocene in age (Feyling-Hanssen 1985; Miller 1985). In a biogeographic sense, the range is similar to that associated with the plant and beetle remains of the Isortoq and Flitaway beds (see above).

(*d*) *Glaciation and oceanography*

The occurrence of subarctic faunas along the Baffin Island coast, north of the present limit, is associated with the extension of the WGC. Conceptually this may be associated with *either* a reduction in strength of the CC or an increase in strength of the WGC, or some combination of both. The palaeoceanographical scenario for the two end-member situations is different, and this difference is critical (see, for example, Andrews *et al.* 1981; Aksu 1981, 1985; Fillon & Aksu 1985; Fillon 1985) to an understanding of the atmosphere–cryosphere–ocean interactions of this region.

Andrews (1972) and Miller (1980) noted that the Holocene raised marine sediments of eastern Baffin Island lack subarctic faunal elements before 8.4 ka BP (9.7 ka BP around outer Frobisher Bay). However, in a very restricted interval, the inshore waters of both Baffin Island and Greenland (Andrews 1972; Hjort & Funder 1974) were invaded by the common blue mussel *M. edulis*, as well as *Mya pseudoarenaria*, *Macoma balthica*, and *Chlamys islandicus*. A similar chronology has been noted in the benthic foraminiferan record in offshore cores (Osterman 1984). *M. edulis* has even been found in the foreset beds of ice-contact glacial marine deltas dated to *ca.* 8 ka BP (Andrews *et al.* 1970). Between *ca.* 5 and 3 ka BP these subarctic elements disappeared from the coast between 66 and 74° N; in the offshore cores, calcareous Foraminifera were replaced by arenaceous forms and (less than 5 ka BP) dissolution of calcium carbonate becomes clear (Osterman 1984). These developments have been associated with either the onset or the strengthening of the Canadian Current, and it is a phenomenon that can be traced for more than 3000 km from the northern Baffin Island shelf, south to the Canadian Maritimes (Scott *et al.* 1984).

The Holocene nearshore marine sequence is thus a cold–warm–cold sequence with the aquatherm being associated with the westward expansion of the WGC. This is associated with an increased advection of warm water in the North Atlantic Drift system. During the same interval, subarctic molluscs expanded along both the East Greenland and Spitsbergen coastlines (Hjort & Funder 1974; Feyling-Hanssen 1955). In the framework of our conceptual stratigraphic model (figure 6), the warm-water influx occurred early in the regressive cycle (that is, during deglaciation) whereas present conditions are marked by cold waters offshore.

Marine conditions during the onset of glaciation have figured prominently in models of glaciation. Two broad and contrasting scenarios have been proposed. In the first, glaciation is associated with relatively warm conditions so as to produce increased accumulation over the surface of the expanding snow–ice surface (Johnson & McClure 1976; Ruddiman & McIntyre 1979; Crowley 1984). In an alternative reconstruction (Denton & Hughes 1983; Denton *et al.* 1986), severe conditions during the onset on glaciation result in the freeze-up of the inner seas and channels of Arctic Canada: in other words, the glacial equilibrium line falls from present values of between 400 and 1200 m (Andrews & Miller 1972) to sea level. This constitutes an extreme form of instantaneous glaciation (Ives *et al.* 1975).

Based on the field collections from the outer coastal forelands of Baffin Island, Andrews (1984) suggested that elements of both models are required to explain the oceanography during ice sheet growth over the northeastern sector of the Laurentide Ice Sheet. In figure 6 a glacial–deglacial cycle is recorded by glacial marine sediments. As noted earlier, these contain marine fossils, preserve terrestrial organic layers (Miller *et al.* 1977; Mode *et al.* 1983) and should preserve the signal of changing nearshore conditions during the transgressive and regressive phases of glaciation. Note that the *major* assumption in this statement is the assertion that the changes in relative sea level along the outer coast are driven by glacial isostasy.

Mode (1985, Figure 17.4) examined the palaeoclimatic indicators in 8 of the 11 aminozones currently delimited on Baffin Island. His survey did not include the poorly studied Loks Land aminozone of middle Wisconsin age, nor the Holocene Eglinton aminozone (figure 7). The assessment of a 'warm' interval is based on the presence of subarctic faunal or floral elements in areas where only arctic taxa occur today. Mode (1985 , p. 515) stated: 'The marine climate was usually warmer than present during the glacial maxima because six of the eight aminozones contain subarctic species *in association with ice-proximal facies*' [italics added]. During the subsequent marine regression, conditions remained warmer than present during four of the eight aminozones. As noted by Mode (1985, p. 515), this was particularly true of the Kogalu aminozone, which contains nine subarctic molluscs in the upper (regressive) units. In addition, terrestrial evidence (*Betula* pollen percentages) also suggest environments warmer than present during this interval.

It is clear from these data that the bulk of fossiliferous raised marine sediments from eastern Baffin Island represent environments characterized by seasonal open-water conditions, terrestrial and marine temperatures warmer than present, and significant glacial meltwater production (formation of large ice-proximal deltas and extensive deposits of silty clays). Enigmatically, evidence for extreme polar climates is lacking. For example, only late Holocene niveo-aeolian sands have been described from the cliff sections. There is virtually no evidence for ice-wedge or sand-wedge casts, although these occur on the modern surface, and no-one has noted ventifacted clasts. It is possible that the vast bulk of such sediments were removed and are represented by the U (unconformities) facies (figure 6).

In the context of these findings, an important question is the age of the various aminozones. Where in the global glacial–interglacial cycles do the aminozones of eastern Baffin Island fall? (figure 9). Do they all represent the same interval within such a hypothetical cycle? Thus the chronology of the aminozones is of vital interest, but, unfortunately, one where appropriate numerical dating methods are not available to provide a solution (see, for example, Szabo *et al.* 1981).

Andrews *et al.* (1985) compared marine and terrestrial events during marine isotope stages 5 and 4 from Baffin Island, West Greenland and Baffin Bay, and concluded that the Kogalu aminozone spans the latter part of stage 5. Thus the development of the last glaciation across northeastern Arctic Canada, the Foxe Glaciation, commenced during this period. However, the Greenland Ice Sheet did not expand and advance onto the shelf until isotopic stage 3 (Kelly 1985).

When using the modern environment as a key to the past, it is important to note that the marine and terrestrial conditions of eastern Baffin Island are controlled by the presence of the Canadian Current. What would result if the channels in the High Canadian Arctic froze over and thickened to form interchannel ice shelves? The Canadian Current would be blocked from Baffin Bay and the oceanographic conditions within the bay would be controlled by the presence and vigour of the West Greenland Current.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

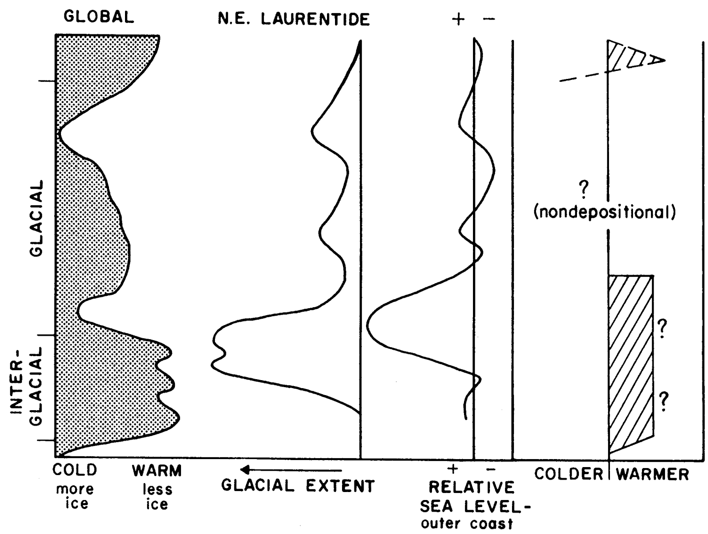

FIGURE 9. Figure 6 represents the glacial–sea-level cycle noted in the forelands (figure 4) whereas this figure portrays the history within a global interglacial–glacial cycle. In a global sense the advance of the northeastern sector on the North American ice sheet may have occurred early in the global glacial cycle. The small advances and sea-level rises associated with advances from a restricted ice sheet may explain low-level sea-level indicators, such as number 2 on figure 4. The (far) right-hand column represents climatic conditions as inferred from the palaeontology of the raised marine units.

CONCLUSIONS

There are significant differences in the interpretation of the late Quaternary palaeoceanography of the northeast and northwest arms of the Atlantic Ocean (Aksu 1985; Aksu & Mudie 1985; Fillon & Duplessy 1982; Fillon 1985; Fillon & Aksu 1985; Kellogg 1975, 1976). It is not clear whether these views represent competing models (i.e. one is right and the other is wrong), or whether both truly represent some version of reality (figure 10). The palaeoceanographic model for the northwest North Atlantic finds support from the inshore palaeontological record (Mode 1985; Andrews *et al.* 1981) and I suggest that glaciation of the northeast sector of the Laurentide Ice Sheet occurred early in a glacial cycle (figure 9). If the information from the sediment sequences, the palaeontology, and associated oceanography is pieced together, it must be concluded that the periods of major global glaciation (e.g. stage 2) are not represented in the Baffin Island raised marine record. This is because extensive regional glaciation occurs at the onset of the cycle, and glacial extent decreases progressively through the remainder of the global glacial cycle (figure 9).

If the relation of regional to global glaciation is as depicted in figure 9, it is tempting to correlate the aminozones with the marine isotopic record. Obviously at this stage, however, and with the dating problems noted earlier, such a correlation is only a hypothesis that requires rigorous testing. Possible tests include the application of new methods, such as thermoluminescence dating, and determining the number of detrital carbonate beds in site 105. On the basis of piston cores, Andrews *et al.* (1985) earlier suggested that the Cape Broughton aminozone is *ca.* 200 ka old.

Kogalu aminozone	stage 4–5
Cape Broughton aminozone	stage 6–7
Kuvinilk aminozone	stage 8–9
Cape Christian aminozone	stage 10–11.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

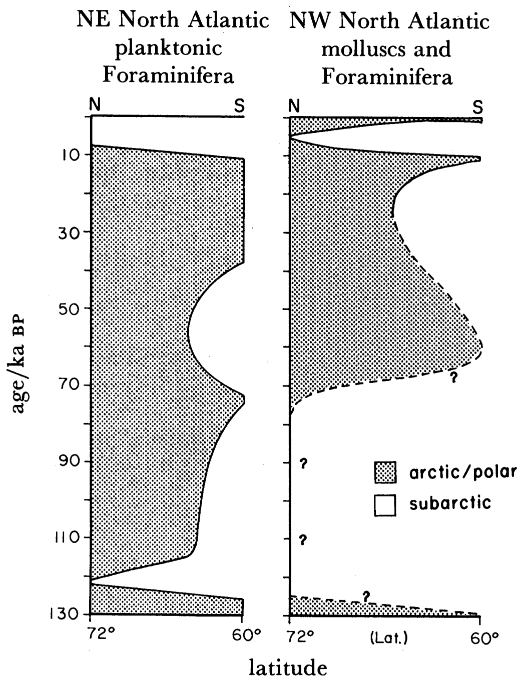

FIGURE 10. Diagrammatic comparison of changes in arctic–subarctic (polar–subpolar?) faunas and floras from the northwestern North Atlantic with those from the Greenland and Norwegian Seas (after Kellogg 1975; Fillon & Aksu 1985; Aksu 1985). Note that these diagrams are based on different biological indicators.

Such a straightforward correlation could be incorrect if there were major changes in the intensity of glaciation in different isotope stages. However, Miller (1985) noted that his age estimates (figure 7) had an 100 ka cyclicity.

The landscape of the eastern Canadian Arctic contains several elements that are rare or unique in heavily glaciated areas. These elements include: the presence of extensive sedimentary forelands that contain packages of Quaternary sediment (figures 4 and 5); the absence of coast-parallel troughs on the shelf; thick sedimentary wedges in the fiords that may pre-date the last glaciation; and the preservation of older non-glacial units on the surface and in valleys of the interior. All these elements together imply that the style of glaciation has been different from that which affected the coasts of West Greenland, Labrador or western Norway.

Marine and terrestrial fossils collected from sequences of raised marine sediments (figures 5–7) normally record intervals when the nearshore environment was warmer than present (i.e. figures 7 and 10). This situation indicates that glaciation of the northeastern sector of the Laurentide Ice Sheet was associated with a reduction in the strength of the Canadian Current and the expansion of the West Greenland Current (figures 9 and 10). Cold, full-glacial global conditions appear to be non-depositional (or erosional) events.

For the past 16 years the research reported in this paper has been supported by the National Science Foundation. I owe a considerable debt of gratitude to the many colleagues quoted in this paper for their contributions to an understanding of the Quaternary history of this pivotal region. In the context of this contribution especial thanks are owed to G. H. Miller, A. R. Nelson, W. N. Mode, A. S. Dyke and J. K. Brigham-Grette.

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Discussion

H. OSMASTON (*Department of Geography, University of Bristol, U.K.*). We enjoyed Professor Andrews' alternative anthropomorphic explanation of the increasing depauperization of the sub-Arctic fauna and flora in successive interglacials (that the species became increasingly browned off with the effort of returning after successive glaciations). However, would he be happy if I paraphrased it thus: that deterministic models, which predict equal returns of species after equal changes of climate, should be replaced by stochastic models, which embody for example the chance elimination of species in biological refugia?

J. T. ANDREWS. In my view, Dr Osmaston's assessment of the situation is correct, or at the very least is a viable alternative to the strictly deterministic association of plants and climate. In the specific case of the Baffin Island data we have to remember that migration of plants to this 500000 km^2 island involves migration across substantial water (in summer) barriers.

R. MARRIS (12 *East Terrace, Budleigh Salterton, Devon, U.K.*). Is it not better to regard the site in Northern Greenland as in a part of a stage in the Beaufort series of deposits until proved otherwise? The nearest on Ellesmere Island are at Yelverton Inlet and south of Alert.

J. T. ANDREWS. Yes, it is possible that the northern Greenland flora at Kapp Kobenhavn is an eastern outlier of the Beaufort Formation of Arctic Canada. However, the suggested age is considerably younger than the accepted Miocene age for the Beaufort. There are forest elements on Banks Island which are younger than the Beaufort and the North Greenland deposits may be correlative with those units.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

FIGURE 4. Oblique air photograph looking northward along the outer coast of eastern Baffin Island. In the foreground is the Qivitu foreland. Numbers 1, 4 and 5 point to specific wave-cut sections: 2 locates a small raised delta some distance below the local marine limit; and 3 points to the right lateral moraine of a former ice lobe moving seaward down Narpaing Fiord.

Figure 5. Photograph of the exposed section at number 5 (figure 4), showing the freshly eroded face of the wave-eroded section.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES