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UPPER CENOZOIC HISTORY OF THE LABRADOR SEA, BAFFIN BAY, AND THE ARCTIC OCEAN: A PALEOCLIMATIC AND PALEOCEANOGRAPHIC SUMMARY

A.E. Aksu,[1] P.J. Mudie,[2] S.A. Macko,[1] and
A. de Vernal[3]

Abstract. Foraminifera, palynomorphs, and stable isotopes in planktonic foraminifera have been studied in piston cores from the northern Labrador Sea, southern Baffin Bay, and the Alpha Ridge region of the central Arctic Ocean. Foraminifera in cores from the Labrador Sea and southern Baffin Bay show glacial-interglacial $\delta^{18}O$ values similar to those reported from the Norwegian and Greenland seas. In contrast to the eastern North Atlantic Ocean, however, peaks of Subarctic foraminifera and dinoflagellates in the western North Atlantic cores show that relatively warm Atlantic surface water continued to flow into the northern Labrador Sea during the early glaciation growth phases of isotopic stages 2, 4, 6, and 8. Boreal and Subarctic pollen suggest that warm Atlantic air flowed into the northern Labrador Sea and Baffin Bay during the ice sheet growth phases. The advection of warm air masses and the presence of open waters during the summer probably provided moisture for the growth of the Laurentide, Innuitian, and Greenland ice sheets. Isotopic and microfossil records in Arctic cores can also be tentatively correlated with major high latitude north Atlantic glacial-interglacial events. Magnetostratigraphic, palynological and amino acid dating, however, shows that sedimentation on the Alpha Ridge has been very slow during the past 0.73 m.y.; therefore, events of less than 20,000 years duration cannot be clearly discerned. Interpretation of pre-Pleistocene paleoenvironments is further limited by the sparseness of calcareous and siliceous microfossils.

INTRODUCTION

Labrador Sea and Baffin Bay constitute a narrow seaway linking the northwest Atlantic Ocean with the Arctic Ocean. This area is situated adjacent to three of the major northern hemisphere ice sheets (Laurentide, Innuitian, and Greenland). The seaway must have played an important role in the transport of heat and moisture between the atmosphere

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[1] Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland, Canada.

[2] Atlantic Geoscience Centre, Geological Survey of Canada, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada.

[3] Département des Sciences de la Terre, Université du Québec à Montréal, Québec, Canada.

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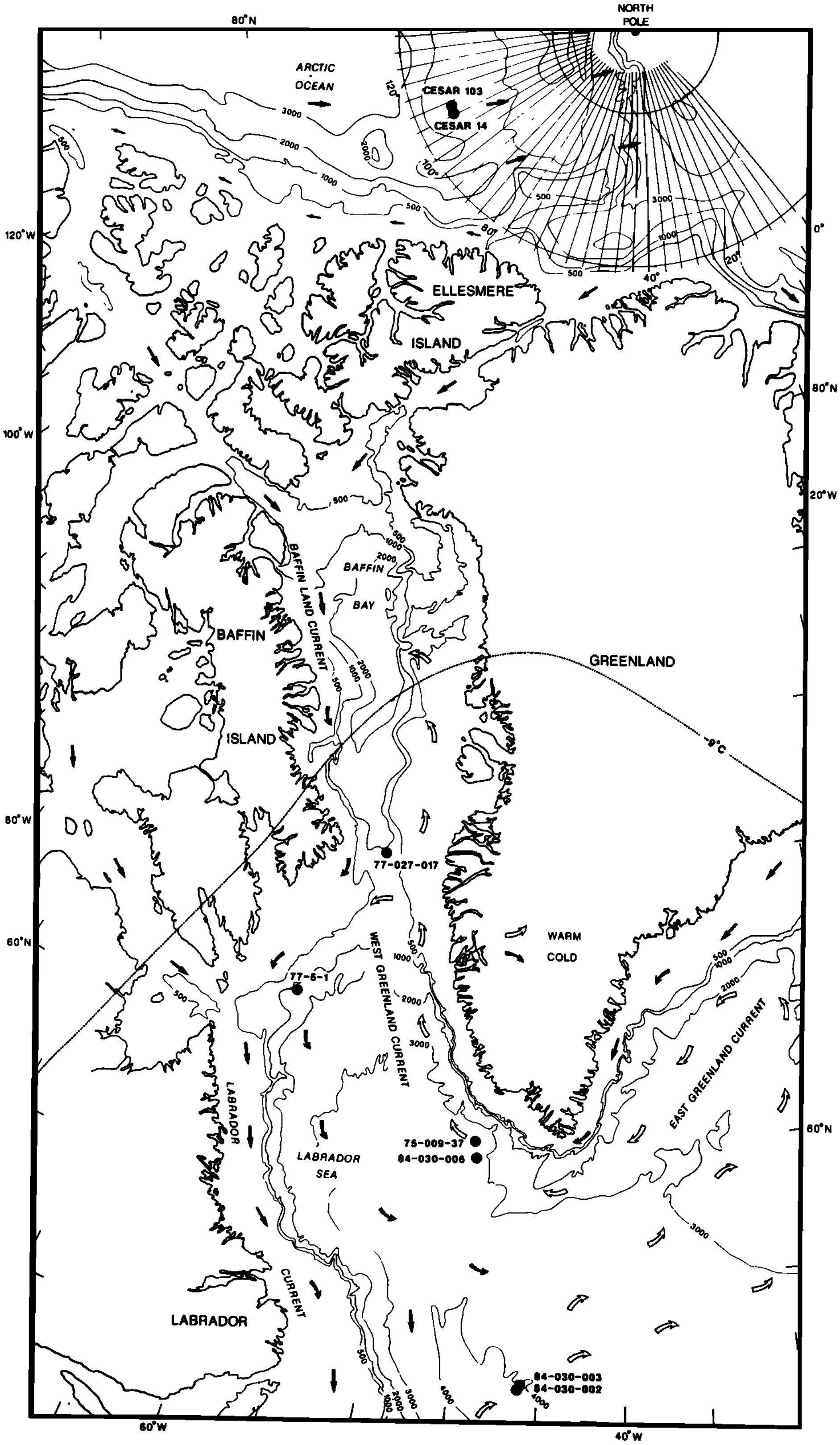
and ocean during the growth and decay of the late Cenozoic glaciation. Previous studies in the Norwegian and Greenland seas and the North Atlantic Ocean suggest that during a transition from interglacial to glacial periods there was a long buildup (approximately 100,000 years) of ice sheets, consisting of multiple rapid ice advances, accompanied by a southerly migration of the Polar Front, which produced extensive year round sea ice cover in these regions [Kellogg, 1976, 1977; Kellogg et al., 1978; Ruddiman et al., 1980]. This North Atlantic model also predicted that perennial sea ice covered the Labrador Sea and Baffin Bay during the glacial intervals. Data from numerous North Atlantic cores indicated that the transitions from glacial to interglacial periods were triggered by increased summer insolation maxima which resulted in the rapid decay of ice sheets [Ruddiman and McIntyre, 1981].

Studies of raised marine deposits in Baffin Island and Greenland, however, indicated that the local glacial buildups were rapid (approximately 15,000 years) and occurred predominantly during the later parts of interglacials which were marked by northward penetration of Subarctic waters into northern Labrador Sea and Baffin Bay [Andrews et al., 1983]. This Arctic model also suggests that ice sheet growth was terminated during the glacial periods due to calving into relatively warm open water. Likewise, it has long been argued that the Arctic Ocean played a major role in regulating the timing of northern hemisphere glaciations, global heat exchange, and oceanic circulation [Clark, 1982; Herman, 1983; Clark et al., 1984; Morris and Clark, 1986]. Detailed interpretation of the upper Quaternary history in the Arctic Ocean has been hampered by the inability to obtain a 1,000 to 2,000 year resolution on core data due to extremely slow sedimentation rates (approximately 1 mm per 1000 years) and the absence of a detailed and reliable chronostratigraphic framework for sediments deposited during the late Neogene and early Pleistocene.

During the past 15 years, lithological and biostratigraphic studies have been made of the late Cenozoic sediments on about 60 cores collected during cruises of Canadian Scientific Ship (CSS) Hudson in the Labrador Sea and Baffin Bay [Fillon and Duplessy, 1980; Aksu, 1981; Aksu and Mudie, 1985a; Fillon and Aksu, 1985; de Vernal, 1986]. In 1983, this data set was expanded by the collection of 26 cores from the Canadian Expedition to Study the Alpha Ridge (CESAR) in the central Arctic Ocean [Mudie and Blasco, 1985; Mudie and Jackson, 1985; Aksu and Mudie, 1985b]. Studies in the southern Baffin Bay [Mudie and Aksu, 1984; Mudie and Short; 1985] and the northern Labrador Sea [Aksu and Mudie, 1985a; Fillon and Aksu, 1985] showed that the micropaleontological and stable isotopic data from several cores cannot be fully explained by either the Atlantic or Arctic model of atmosphere-hydrosphere-cryosphere interactions during late Pleistocene the glacial-interglacial cycles. A modification of the Atlantic model was therefore proposed [Aksu and Mudie, 1985a] which has now been tested further with data from the Southern Labrador Sea, as presented in this paper. At the same time, further progress has been made in resolving the chronology and biostratigraphy of the Arctic cores. A study on amino acid epimerization of planktonic foraminifera in the Arctic Ocean [Macko and Aksu, 1986], confirms the slow sedimentation rate for the Alpha Ridge and shows that a relatively detailed time framework can be established for the upper Cenozoic sediments.

The purpose of this paper is to review the paleoclimatic and paleoceanographic data available for sites along a transect from 53° to 67°N in the Labrador Sea and from adjacent parts of the western Arctic Ocean (Figure 1) and to place this information in a broader context of an ocean-atmosphere circulation model. Planktonic microfossil and oxygen isotopic data have been collected for each of the sites shown in Figure 1, and these data are supplemented by studies of organic-walled microfossils, including marine dinoflagellate cysts and terrestrial pollen and spore assemblages. Amino acid epimerization on planktonic

Fig. 1. (Opposite) Generalized surface circulation map of the Labrador Sea, Baffin Bay, and the Arctic Ocean showing locations of cores used in this study. Isobaths are in meters. The -9°C mean annual air isotherm marks the average position of the Arctic-Atlantic air front.


NORTH POLE
80°N
ARCTIC OCEAN
3000
2000
1000
500
120°
CESAR 103
CESAR 14
100°
80°
500
3000
1000
0°
20°
40°
120°W
ELLESMERE ISLAND
80°N
100°W
20°W
500
1000
2000
BAFFIN BAY
BAFFIN LAND CURRENT
BAFFIN ISLAND
GREENLAND
-9°C
80°W
77-027-017
60°N
WARM
COLD
77-5-1
WEST GREENLAND CURRENT
EAST GREENLAND CURRENT
LABRADOR CURRENT
LABRADOR SEA
75-009-37
84-030-006
60°N
LABRADOR
84-030-003
84-030-002
4000
3000
60°W
40°W

foraminifera was also studied where needed to refine the chronostratigraphy of the reference cores. Details of the microfossil and palynomorph assemblages are reported in the original papers describing individual sites. The intent of this paper is not to provide an exhaustive review of all data but to provide a state-of-the-art summary for our knowledge up to 1987.

OCEANOGRAPHY

Currents in the Canadian Arctic (Figure 1) reflect the outflow of water from the Arctic Ocean, which, in turn, is the result of the excess inflow into the Arctic Ocean from northern branches of the North Atlantic Current, the North Pacific Current and runoff from large rivers of northern Siberia and Canada. Within the Arctic Ocean, surface water moves towards the North Pole with velocities of 2 to 3 cm s^{-1}, becoming strongest as the return flow southwards forms the East Greenland Current (Figure 1) [Coachman and Aagard, 1974]. The East Greenland Current forms the main outflow of Arctic surface water. Along the Canadian Polar margin, surface water flows southwest with seasonally variable current velocities. A large volume of Arctic water then flows south through channels of the Canadian Arctic Archipelago into Baffin Bay to form the Baffin Land Current. Arctic overflow water also flows through Hudson Strait south of Baffin Island and joins the Baffin Land Current to form the Labrador Current in the western Labrador Sea.

During the summers in the Labrador Sea, the weaker (approximately 2 to 4 cm s^{-1}) and warmer (2° to 6°C) West Greenland Current, consisting of a mixture of East Greenland and Irminger currents, flows northward along the west coast of Greenland into the central Baffin Bay (Figure 1). The cold Baffin Land Current, originating from the Arctic Ocean, flows southward along the Baffin Island coast with current velocities of 2 to 25 cm s^{-1}. A branch of the West Greenland Current merges with the Baffin Land Current south of the Davis Strait. In the winter, the West Greenland Current is displaced southward, and the cold Baffin Land Current dominates most of the northern Labrador Sea [Coachman and Aagaard, 1974].

CORE DATA

Baffin Bay and Davis Strait

The first detailed paleoceanographic studies [Mudie and Aksu, 1984; Fillon and Aksu, 1985; Mudie and Short, 1985; Aksu and Mudie, 1985a] were concentrated on a set of cores from the Davis Strait which separates the small Arctic Baffin Bay basin from the mixed Arctic-Subarctic waters of the Labrador Sea. In the Davis Strait strong mixing occurs between surface water outflowing from the Arctic Ocean and Subarctic water inflowing from the North Atlantic Ocean. Therefore a clear signal of paleotemperature changes accompanying the glacial-interglacial shifts of the Polar Front as described for the central and northeast Atlantic Ocean [Ruddiman and McIntyre, 1981] will occur in this region. The Davis Strait also lies just south of the Arctic/Atlantic air front (Figure 1) and is expected to include an amplified signal of atmospheric circulation changes as recorded by pollen-spore assemblages in the marine sediment cores [Mudie, 1982; Mudie and Short, 1985].

Most cores from central Baffin Bay are unsuitable for high resolution studies because turbidites, redeposited microfossils, and calcite dissolution obscures the paleoecological record [Mudie and Short, 1985; Praeg et al., 1986; de Vernal et al., 1987]. Core 77-027-017 from southern Baffin Bay provides a continuous record of planktonic foraminifera and oxygen isotopic data for the past 0.3 m.y. (Figure 2) [Mudie and Aksu, 1984]. The isotopic curve at this site is similar to those reported for comparable latitudes in the Norwegian Sea [Kellogg, 1977]. However, the Baffin Bay record differs from those of the Norwegian Sea in that it contains peaks of Subarctic foraminifera generally corresponding to early glacial stages 2, 4, 6, 8, and 10 (Figure 2), while the Norwegian Sea records show comparable peaks for the interglacial stages 1 and 5 [Kellogg, 1977]. The Subarctic foraminiferal fauna include Neogloboquadrina pachyderma dextral, Globigerina bulloides and G. quinqueloba. The modern distribution of these species is confined to warmer (4° to 10°C) surface waters of the West Greenland Current. Furthermore, palynological data in core 77-027-017 show that peaks of

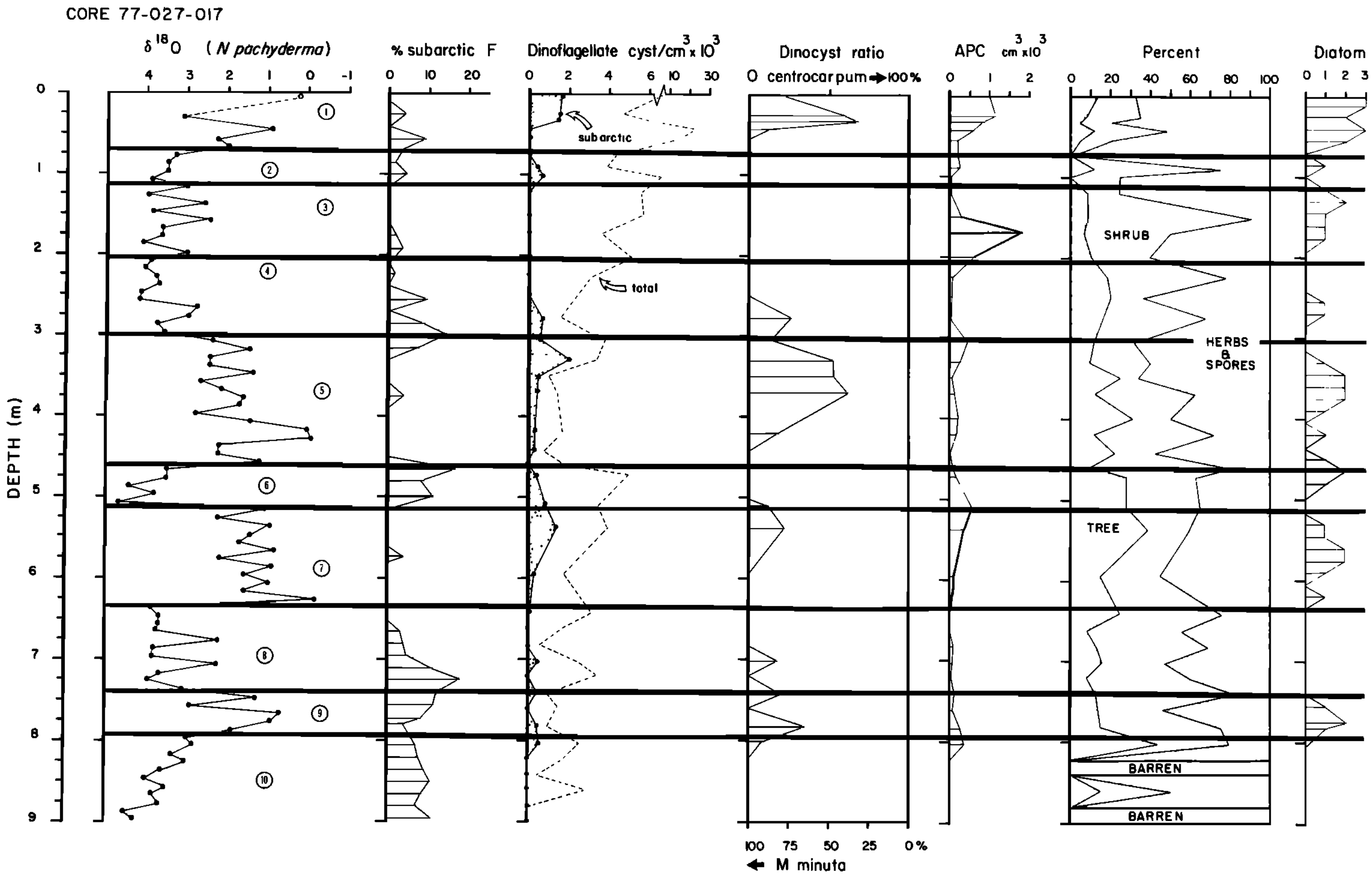


Fig. 2. Summary of micropaleontological data in core 77-027-017, showing the $\delta^{18}O$ record, percentage of total subarctic planktonic foraminifera, dinoflagellates and total diatoms (in >63-micron fraction). The diatom scale is 0 for absent, 1 for rare, 2 for present, and 3 for abundant. APC is absolute pollen concentration, cumulative percentages of conifer tree, shrub (*Alnus* and *Betula*) and herb pollen, and moss spores. Circled numbers refer to oxygen isotopic stages.

subarctic dinoflagellate indicators mainly occur in the interglacial isotopic stages, but they are also present in glacial stage 2 and the lower parts of stages 4, 6, and 8 (Figure 2). The Subarctic dinoflagellates include Operculodinium centrocarpum, Nematosphaeropsis labyrinthea, Bitectadodinium tepikiense, Spiniferites ramosus, S. elongatus, S. bulloideus and S. membranaceus. These species presently have their maximum percentage abundances in the temperate north Atlantic, particularly beneath the North Atlantic Drift [Harland, 1983]. Their present-day distributions in Baffin Bay are confined to deepwater areas beneath the warm West Greenland Current [Mudie and Short, 1985]. The presence of Subarctic foraminifera and dinoflagellates in early glacial stages of core 77-027-017 suggests significant advection of Subarctic surface waters into the Labrador Sea during these times [Aksu, 1981; Mudie and Aksu, 1984; Fillon and Aksu, 1985]. The apparent absence of Subarctic foraminiferal assemblages in the interglacial stages in southern Baffin Bay probably reflects calcium carbonate dissolution which is known to affect the foraminiferal distribution in surface sediments of central Baffin Bay [Aksu, 1983].

The number and diversity of Quaternary pollen and spores in core 77-027-017 are low (Figure 2), as expected for oceanic sediments north of the treeline in eastern Canada [Mudie, 1982; Mudie and Short, 1985]. However, there are cyclical variations in absolute pollen concentrations and percentage abundances of indicator species which can be correlated with oxygen isotopic stages [Mudie and Aksu, 1984]. Lower pollen concentration and diversity during the glacial stages suggest a southerly shift of the Arctic air front and stronger zonal flow of polar air characterized primarily by Betula, Artemisia, and Gramineae. The presence of Alnus, Picea, and Pinus boreal tree and shrub pollen in early glacial isotopic stages 4, 6, and parts of 8 and 10 indicates influx of Atlantic air in all initial and some later phases of continental ice sheet growth [Mudie and Aksu, 1984].

Labrador Sea

In order to determine the regional extent of the Subarctic fauna and flora found in glacial stages of core 77-5-1, a detailed study of calcareous foraminifera, siliceous microfossils (diatoms and radiolarians), and palynomorphs was made on northwestern Labrador Sea cores [Aksu and Mudie, 1985a]. These cores, including core 77-5-1 (Figure 1) can be correlated with cores from the northeastern Labrador Sea [Fillon and Duplessy, 1980; de Vernal, 1986]. Microfossil and isotopic records for core 77-5-1 (Figure 3) show a similar correlation between the Subarctic foraminiferal distribution and glacial isotopic stages as found in Davis Strait core 77-027-017. The salient features of core 77-5-1 can be summarized as follows:

1. Subarctic planktonic foraminifera are present throughout the core except in the upper part of isotopic stage 2 (Figure 3). In general, glacial isotopic stages (particularly the middle and some early parts) include higher percentages of Subarctic foraminifera. Relatively lower total foraminiferal abundances as well as Subarctic planktonic foraminifera in interglacial stages largely reflect intensified dissolution during these intervals [Aksu and Mudie, 1985a].
2. Atlantic dinoflagellate assemblages predominantly occur in interglacial isotopic stages but also occur in the early and middle parts of glacial isotopic stages 2, 4 and 6 and sporadically in stage 8 (Figure 3).
3. As in core 77-027-017 from the Northern Davis Strait, both foraminiferal and dinoflagellate data indicate that Subarctic waters penetrated northward into the northwest Labrador Sea during the early glacial stages.
4. Fluctuations in total radiolarians [Aksu and Mudie, 1985a] broadly correspond to the changes in the Subarctic foraminiferal fauna. In contrast, total diatoms show maxima during interglacial stages, which broadly correspond to peaks in the dinoflagellate flora.
5. Quaternary pollen and spores are present in low concentrations throughout the core, however, percentage frequencies of the most common species, however, show fluctuations which are broadly correlated with the oxygen isotopic stages (Figure 3). Glacial isotopic stages are mainly characterized by the predominance of herb pollen (Gramineae, Artemisia, Rosaceae, and Chenopodiaceae) and Polytrichum-type moss spores, reflecting the dominance of very cold and dry westerly Arctic and

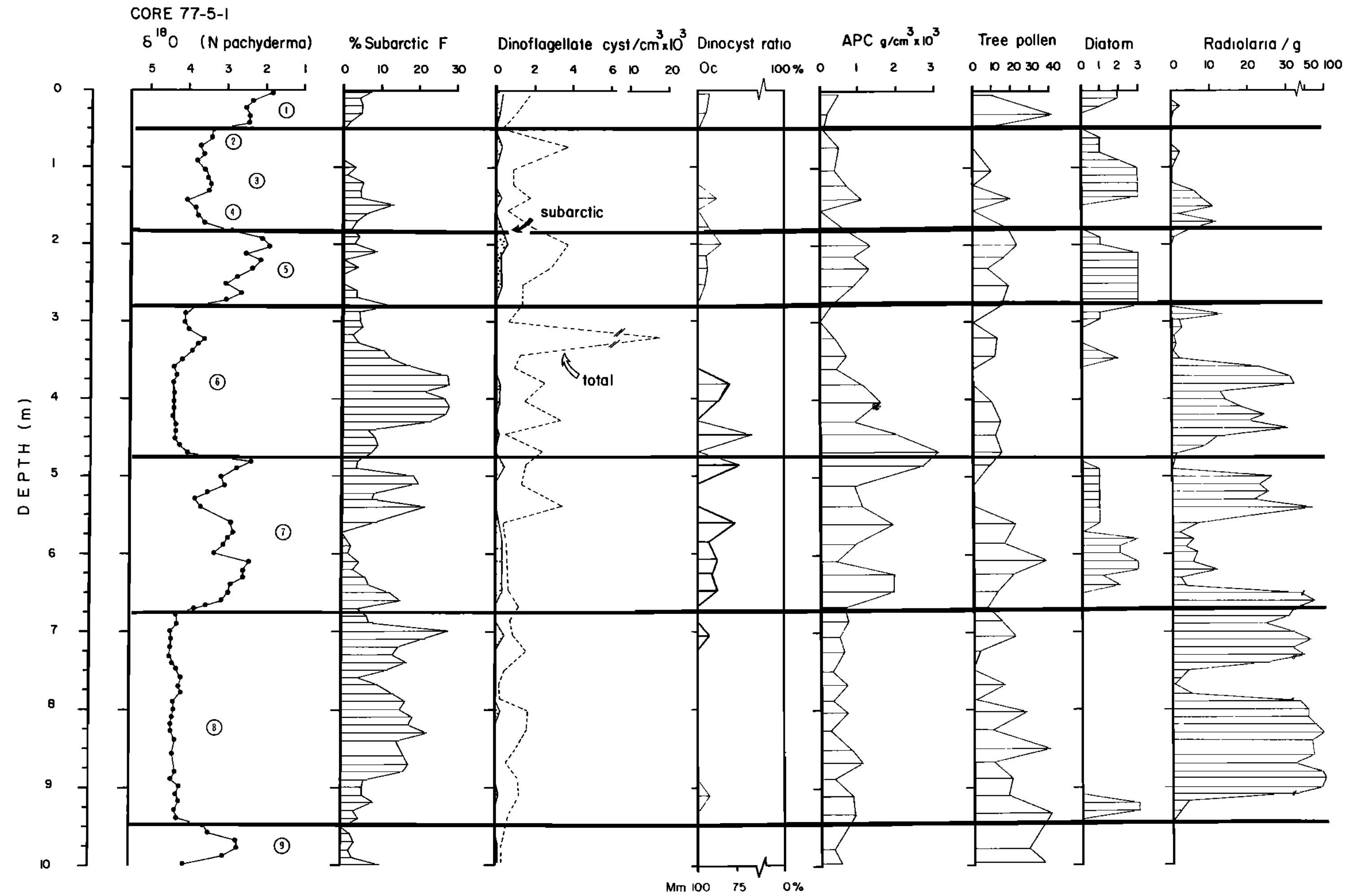


Fig. 3. Summary of micropaleontological data in core 77-5-1, showing the $\delta^{18}O$ record, percentage of total Subarctic planktonic foraminifera and dinoflagellates, total radiolaria, and diatoms (in >63-micron fraction). Diatom scale is 0 for absent, 1 for rare, 2 for present, and 3 for abundant. Circled numbers refer to oxygen isotopic stages. Dinocyst ratio is the percentage ratio of Operculodinium centrocarpum (Oc) to Multispinula minuta (Mm). APC is absolute pollen concentration; tree pollen are given in percentage and include Picea plus Pinus plus Carya.

Arctic Archipelago air masses [Barry and Chorley, 1971]. Higher pollen and spore concentrations and peaks of boreal tree and shrub pollen occur in early glacial isotopic stages. These palynological data suggest northward transport of pollen in southerly (cyclonic) airmasses during the onset of glaciations (Figure 3).

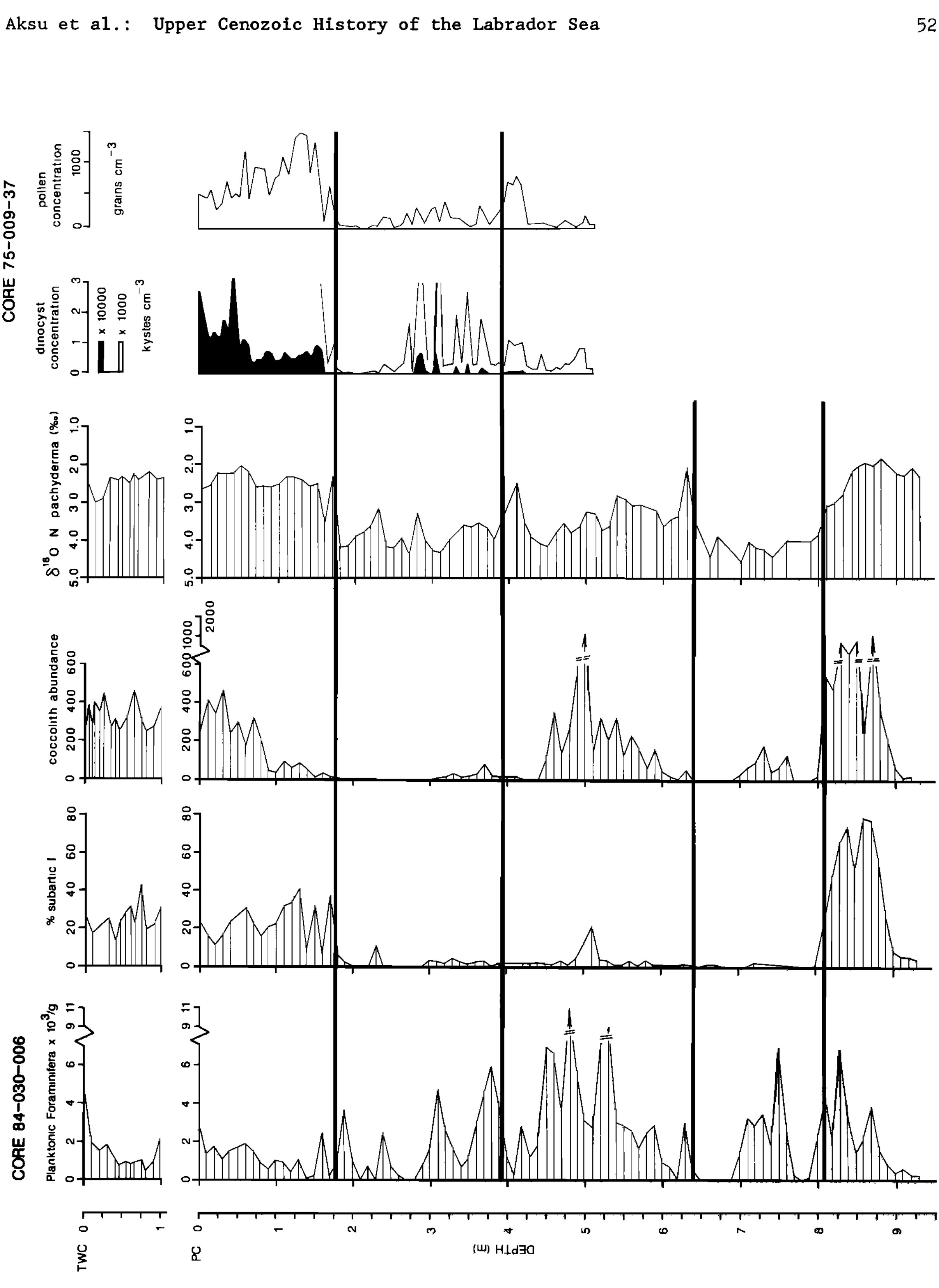
Cores 84-030-006 and 75-009-37 were studied from the southwestern margin of Greenland (Figure 1). These cores are from an area of predominantly hemipelagic deposition located between contourites on the northwestern flank of Eirik Ridge and turbidites of the North Atlantic Mid-Ocean Channel [Aksu et al., 1986; de Vernal, 1986; Scott et al., 1986]. Core 84-030-006 recovered about 9 m of brown and olive grey, moderately bioturbated silty and clayey muds. This core contains an oxygen isotopic record extending into the upper stage 5 with glacial-interglacial changes of about 2 mil (Figure 4) [Aksu et al., 1986; Scott et al., 1986]. Core 75-009-037 penetrated about 8 m into sediments similar to those recovered in core 84-030-006. Oxygen isotopic data obtained for the upper 3 m of the core show that the stage 1/2 transition occurs at 290 cm depth [de Vernal, 1986]. Sedimentation rates based on several ^{14}C dates suggest that the base of this core is younger than about 75,000 years.

Interglacial isotopic stages 1 and 5 in core 84-030-006 are characterized by a generally high abundance of coccoliths, dominated by Emiliania huxleyi and Gephyrocapsa oceanica, and by peaks of Subarctic planktonic foraminifera, dominated by G. quinqueloba and to a lesser extent by G. bulloides, N. pachyderma dextral, Globigerinitida uvula, and Ga. glutinata [Aksu et al., 1986]. These floral and faunal data suggest high primary productivity during interglacial stages and the presence of relatively warmer surface water masses in the eastern Labrador Sea during these times. The lower parts of glacial stages included abundant coccoliths and Subarctic planktonic foraminifera, suggesting the advection of Subarctic waters into the Labrador Sea. The upper parts of glacial isotopic stages 2 and 4, however, are characterized by the absence or very low abundance of coccoliths and Subarctic planktonic foraminifera.

Interglacial isotopic stage 1 in core 75-009-37 is characterized by high dinocyst concentrations dominated by Subarctic dinoflagellates (particularly by O. centrocarpum and N. labyrinthea and to a lesser extent S. ramosus and S.elongatus), indicating high primary productivity in relatively warm surface waters (Figure 4) [de Vernal, 1986]. These sediments also include relatively high concentrations of pollen and spores dominated by Picea and Pinus and to a lesser extent by Betula, Alnus, Lycopodium, and Sphagnum, representing influxes of southern air masses similar to those occurring during present times. Glacial isotopic stage 2 is characterized by very low pollen, spores, and dinoflagellate concentrations. As in the northwestern Labrador Sea cores, the lower part of this interval is also marked by sporadic occurrences of Subarctic dinoflagellate assemblages (mainly O. centrocarpum and Brigantedinium simplex).

Cores 84-030-002 and 84-030-003 were studied from the southern Labrador Sea (Figure 1), in an area of pelagic sediments between the turbidites of the North Atlantic Mid-Ocean Channel and the drift deposits covering the southwest Greenland continental rise [Scott et al, 1986; de Vernal, 1986]. These cores contain alternating sequences of brown muddy carbonate ooze and grey silty or clayey muds similar to lithologies found in late Cenozoic sediments south of the Grand Banks [Alam et al., 1983). A composite stratigraphy for the southern Labrador Sea cores (Figure 5) shows a sequence of oxygen isotopic events of similar magnitude to those marking stages 1 to 8 in Norwegian Sea [Kellogg et al., 1978]. The palynostratigraphy for core

Fig. 4. (Opposite) Summary of micropaleontological data in cores 84-030-006 and 75-009-37, showing the $\delta^{18}O$ record, Subarctic planktonic foraminifera and coccolith abundances (in <10-micron fraction) in 84-030-006 and dinoflagellate, pollen, and spore concentrations in 75-009-37. Core 75-009-37 is replotted taking the isotopic stage 1/2 boundary as a reference datum to allow comparison of micropaleontological data in these cores. Coccolith abundance is calculated on the basis of the average of 100 SEM frames. TWC is trigger weight core.


CORE 75-009-37
pollen concentration
0
1000
grains cm^{-3}
dinocyst concentration
0
1
2
3
x 10000
x 1000
kystes cm^{-3}
δ^{18}O N pachyderma (‰)
1.0
2.0
3.0
4.0
5.0
coccolith abundance
0
200
400
600
1000
2000
% subarctic I
0
20
40
60
80
CORE 84-030-006
Planktonic Foraminifera x 10^3/g
0
2
4
6
9
11
TWC
0
1
PC
0
1
2
3
4
5
6
7
8
9
DEPTH (m)

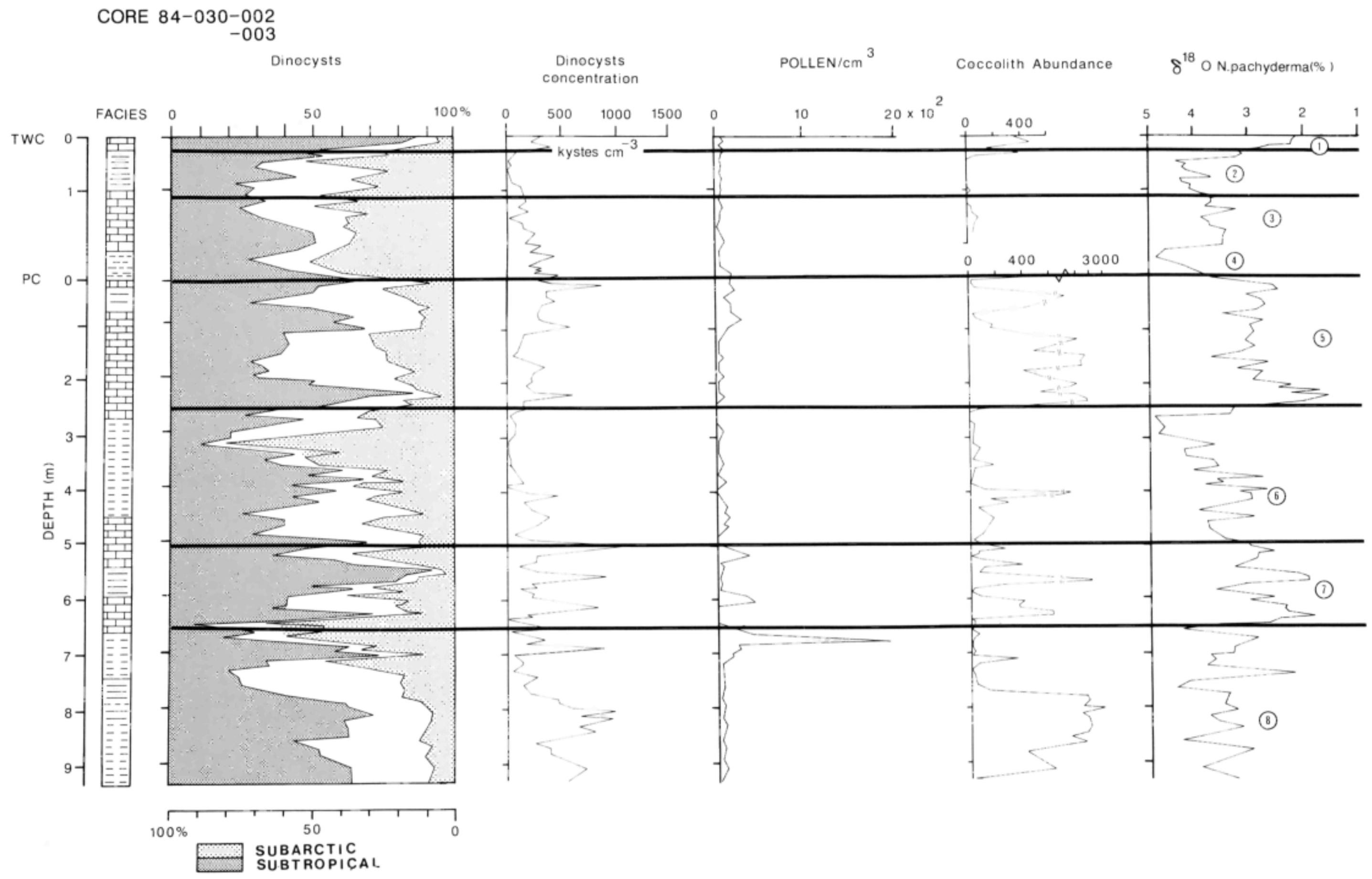


Fig. 5. Composite summary of micropaleontological data in cores 84-030-002 and 84-030-003, showing the δ^{18}O record, dinocyst assemblages, dinoflagellate and pollen concentrations, and coccolith abundances (in <10-micron fraction). Coccolith abundance is calculated on the basis of the average of 100 SEM frames. TWC is trigger weight core.

84-030-003 is described in detail elsewhere [de Vernal and Hillaire-Marcel, 1987] and is summarized as follows:

1. Both Subarctic and Subtropical dinocyst species are present throughout isotopic stages 1 to 8 (Figure 5). Subarctic dinoflagellates are similar to those found in northern Labrador Sea cores; Subtropical dinoflagellates include mainly *Impagidinium* spp. (*I. patulum*, *I. aculeatum*, *I. strialatum*, *I. paradoxum*, and *I. sphaericum*). These floral data indicate continuous flow of warm surface water into southern Labrador Sea during the past approximate 0.3 m.y.
2. Major changes in total numbers of dinocysts per cubic centimeter show fluctuations that generally indicate reduced phytoplankton productivity during glacial intervals, although productivity remains relatively high during early parts of stages 6 and 8.
3. Total numbers of *O. centrocarpum*, *N. labyrinthea*, *Spiniferites* spp., and *Bitectadodinium tepikiense* (the main Subarctic water mass indicators in Davis Strait and southern Baffin Bay cores) have peaks in early parts of stages 4, 6, and 8, showing good correlation with dinoflagellate records of cores 77-027-017, 77-5-1, and 75-009-37.
4. Pollen and spore concentrations are relatively low throughout the southern Labrador Sea cores, as expected from the distance offshore and the pelagic nature of the sediments [Mudie, 1982; 1986]. Peaks of boreal tree pollen (mostly *Picea*) mark the upper parts of interglacial stages 5 and 7; however, significantly large numbers of *Picea* pollen persist into early parts of glacial stages 4 and 6 (Figure 5). This pattern of tree pollen influx is generally similar to that found in Davis Strait cores (Figure 2). In core 84-030-003, a large peak in pollen concentration at the top of isotopic stage 8 reflects increases of both *Picea* tree pollen and *Artemisia* shrub/herb pollen, which are also evident in the pollen diagram for core 77-5-1 from Davis Strait [Aksu and Mudie, 1985a].

The coccolith stratigraphy for core 84-030-003 is described elsewhere [Hearn, 1986]. The important features are summarized as follows:

1. Except for two barren intervals, coccoliths are present throughout the core in varying abundances (Figure 5). Subarctic coccoliths (mainly *Gephyrocapsa oceanica*) are present in isotopic stages 1 through 8, confirming the presence of warmer surface waters in the core site during the last 0.3 m.y. as suggested by the dinoflagellate data.
2. Interglacial isotopic stages 1, 5, and 7 are generally characterized by relatively high coccolith abundances. The isotopic stage 6/5 transition is marked by a conspicuous absence of coccoliths, whereas the 2/1 and 8/7 transitions are marked by very low coccolith abundances, probably indicating considerable surface water salinity reductions in the Labrador Sea associated with major deglaciations.
3. Glacial isotopic stages 4, 6, and the upper part of 8 include relatively lower coccolith abundances compared to the early parts of isotopic stages 8 and 6, which have much higher coccolith abundances. High coccolith abundances are indicative of relatively high phytoplankton productivity during these intervals as suggested also by the dinoflagellate data.
4. This pattern of coccolith distribution is similar to that found in the eastern Labrador Sea core 84-030-006 [Aksu et al., 1986].

Arctic Ocean

Detailed chronostratigraphic, biostratigraphic, and paleoenvironmental studies have been carried out on three cores from the southeastern Alpha Ridge; CESAR 102, 103, and 14. Palynomorphs in CESAR cores 14 and 103 were studied in the same samples used for paleomagnetic measurements [Mudie, 1985; Aksu and Mudie, 1985b]. It was found that the cores contained the range tops and/or bases (Figure 6) of several dinocyst taxa with well-constrained ages based on calibration against magnetochronological and biochronological datums at Deep Sea Drilling Project (DSDP) and Ocean Drilling Project (ODP) high latitude drill sites in the north Atlantic and Pacific oceans. These ranges appear to confirm the validity of the magnetochronology proposed for the CESAR cores [Aksu, 1985b]. Several pollen genera (e.g., *Liquidambar*, *Podocarpus*, and *Nyssa*), however, also show last appearance datums which are more difficult to interpret, because subtropical and temperate tree genera were eliminated from the Canadian Arctic

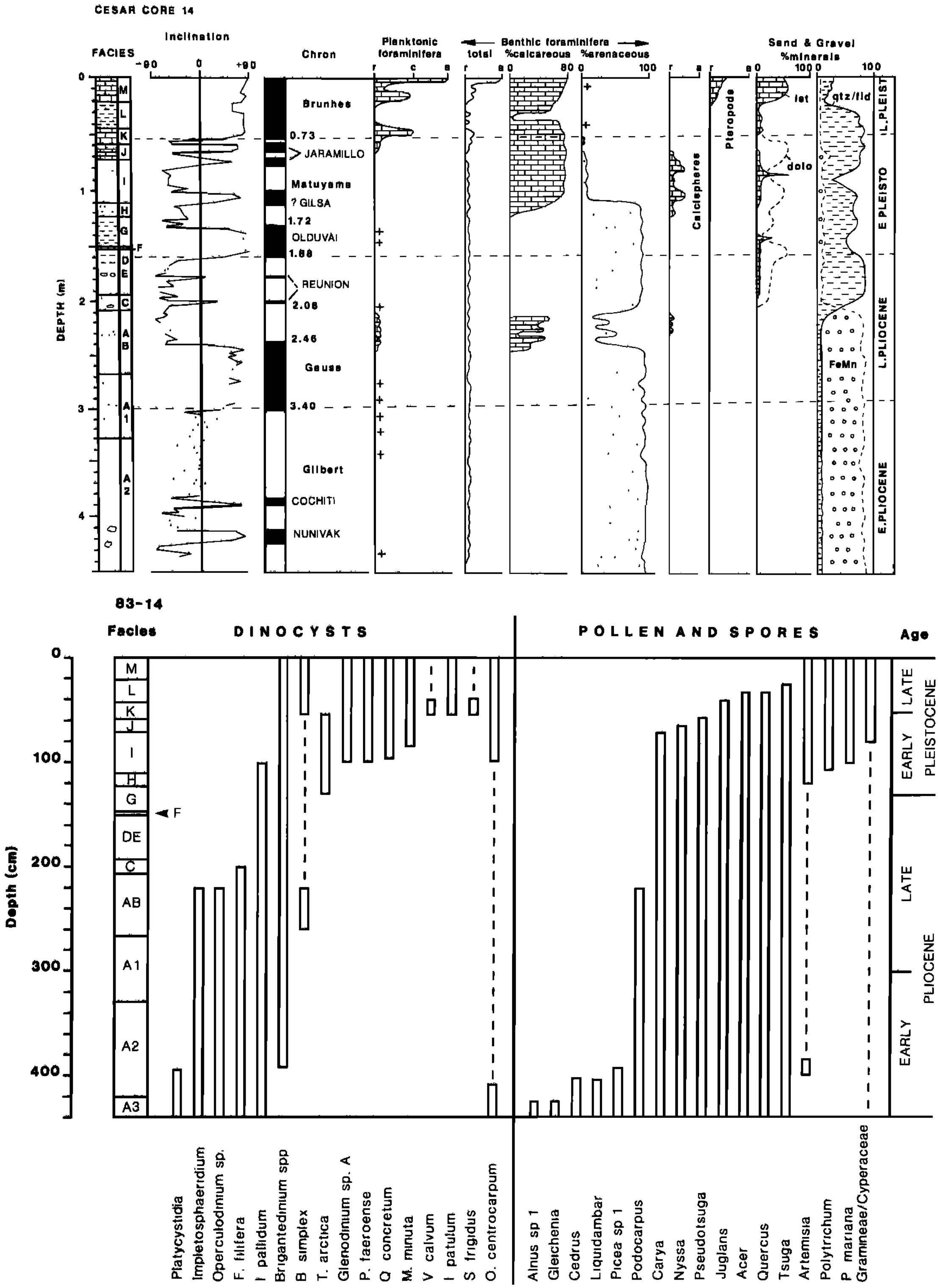


Fig. 6. Summary of magnetostratigraphic and micropaleontological data in core CESAR 14. A2 to M are lithofacies discussed in Mudie and Blasco [1985]. The microfossil abundance scale is r for rare, c for common, and a for abundant.

during the Miocene, although they persisted into the Pliocene or middle Pleistocene in northwest Europe or northern Asia [Leopold, 1969].

The pollen stratigraphy therefore raised the question of whether the terrestrial palynomorphs in the Alpha Ridge cores were largely reworked or whether the tree pollen reflected long-distance transport in air masses originating from the eastern Arctic. For the Pleistocene interval, this question has now been clarified by studies of airborne pollen trajectories in the Canadian high Arctic [Bourgeois et al., 1985; Mudie and Matthiessen, unpubl. data]. Regional and seasonal distribution of pollen in snow samples from the Queen Elizabeth Islands and Arctic Ocean ice cover indicate extremely long trajectories with strong zonal components. Synoptic air circulation data suggest that tree pollen are mainly transported to the central Arctic Ocean by high-latitude cyclonic storm tracks that move across Europe and Siberia during the summer, with smaller amounts transported by the low-altitude anticyclonic circulation that prevails during most of the year. The chronostratigraphic value of pollen in the pre-Pleistocene sediments, of course, depends on the assumption that mean air circulation was essentially the same as during the Pleistocene. This assumption is generally supported by data from ODP Site 646 in Labrador Sea [de Vernal and Mudie, 1987].

The calcareous microfossil records of the CESAR cores include several important features:

1. Planktonic foraminifera are predominant only in the upper Pleistocene sediments in long piston cores from the Alpha Ridge (Figure 6). In shorter gravity cores (Figure 7), where undisturbed surface sediments are recovered, planktonic foraminifera show large variations in abundance, ranging from 10 to 10,000 specimens per gram sediment. Planktonic foraminifera also occur in much lower abundances in a short interval in the late Pliocene (Figure 6) [Scott et al., 1987].
2. Calcareous benthic foraminifera show a similar pattern of distribution to that of the planktonic foraminifera, but they are more abundant in the lower Pleistocene sediments, starting above the Olduvai magnetic subchron (Figure 6).
3. A major benthic faunal turnover occurs in the earliest Pleistocene where there is a change from 100% arenaceous benthic foraminifera to more than 80% calcareous benthic foraminifera, associated with a sharp increase in the occurrence of the calcisphere, Thoracosphaera arctica [Gilbert and Clark, 1982]. A faunal turnover of short duration is also found in the late Pliocene, immediately above the Gauss magnetochron where the exclusively arenaceous fauna is temporarily replaced by a predominantly calcareous fauna, followed by a return to exclusively arenaceous benthic foraminiferal fauna (Figure 6) [Scott et al., 1987].
4. Calcareous-walled dinoflagellates [calcispheres of Gilbert and Clark, 1982] only occur in the early Pleistocene sediments, between the Olduvai subchron and the Brunhes Matuyama boundary (Figure 6).
5. Pteropods occur only in the upper approximately 20 cm of late Pleistocene sediments.

Planktonic foraminifera and oxygen isotopes have been studied in two short gravity cores (CESAR 102 and 103) from the eastern Alpha Ridge area of the Arctic Ocean [Aksu, 1985a; Scott et al., 1987]. These cores include down-core sequences of sediments similar to those of other western Arctic Ocean cores [Clark et al., 1980; Mudie and Blasco, 1985]. Paleomagnetic data from both cores [Aksu, 1985b] and amino acid epimerization in planktonic foraminifera in core 102 [Macko and Aksu, 1986] indicate that the average sedimentation rate in the area is about 1 mm per 1000 years (Figure 7). In both cores, the oxygen isotopic data of N. pachyderma (sinistral) exhibit a record where classic oxygen isotopic stages are not readily identifiable (Figure 7). The data show large fluctuations in oxygen isotopes up to 2.5 mil; however, the amplitude of 18-0/16-0 in the isotopically light intervals also varies from 1.4 to 2.4 mil. Detailed examination of X-radiographs from both cores show sedimentary structures typical of bioturbation. Slow sedimentation rates and bioturbational mixing thus probably reduce the time resolution and amplitude of the oxygen isotopic record. On the basis of calculated sedimentation rates [Aksu and Mudie, 1985b; Macko and Aksu, 1986], a first-order estimate is

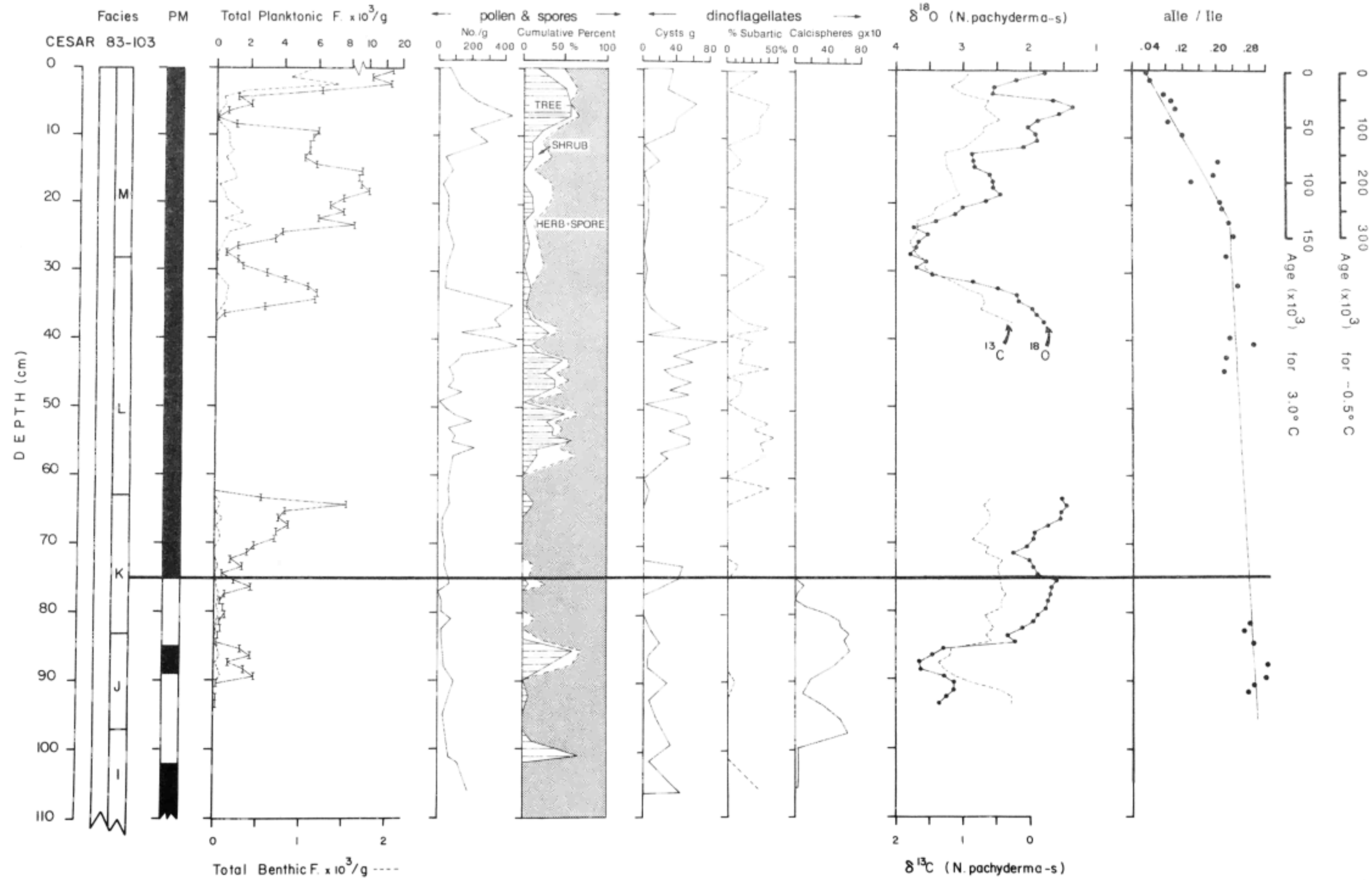


Fig. 7. Summary of micropaleontological and stable isotopic data in CESAR core 103 plotted together with paleomagnetic time scale and averaged epimerization of isoleucine in *N. pachyderma* sinistral (from CESAR 102). I to M are lithofacies discussed in Mudie and Blasco [1985].

made for the ages of isotopically light and heavy intervals (Figure 7). Differences between the isotopic signals for the Arctic and North Atlantic indicate that the oxygen isotopic fluctuations in the CESAR cores are controlled primarily by glacial-interglacial changes in the isotopic composition of the Arctic Ocean surface waters, which mainly reflect changing patterns of freshwater input by large rivers from Siberia and northern Canada and, to a lesser extent, changes in the continental ice volume and Atlantic inflow from the Norwegian-Greenland seas [Aksu, 1985a].

In CESAR cores 102 and 103, N. pachyderma (sinistral) constitute more than 80% of the planktonic foraminiferal assemblage, with N. pachyderma dextral, G. quinqueloba, and G. egelida being secondary in importance [Aksu, 1985a]. Dinoflagellate cyst concentrations are generally very low in most of the Pleistocence sediments, which is typical for perennially ice-covered waters [Mudie, 1985]. Pollen and spore concentrations are also low, as expected in deep-sea Arctic Ocean sediments. In the upper approximately 55 cm of the core, Pinus and Picea constitute about 50% of the total pollen, with sporadic and rare occurrences of Abies, Tsuga, Acer, and Quercus [Mudie, 1985]. These tree pollens reflect long distance transport from boreal forests. Gramineae, Cyperacea, Rumex, and Artemisia are the most common herb pollen and are probably of Arctic origin. Betulanana and Salix shrub pollen are probably of circumpolar origin [Mudie, 1985].

Downcore foraminiferal, dinoflagellate, and pollen and spore concentrations exhibit large variations in abundance (Figure 7), probably reflecting the superimposed effects of changes in productivity controlled primarily by the extent and thickness of sea ice and variations in surface water salinity due to changes in river runoff volume [Aksu, 1985a; Mudie, 1985]. Between 4 and 10 cm in core 103, foraminiferal abundances are very low, and oxygen isotopic values are extremely depleted (Figure 7). This interval also includes relatively high concentrations of dinocysts, pollen, and spores, with tree pollen (mostly Picea and Pinus) constituting approximately 50% of the pollen and spore counts. On the basis of constant sedimentation rates of about 1 mm per 1000 years [Aksu and Mudie, 1985b; Macko and Aksu, 1986] the ages for the upper and lower boundaries of this light isotopic interval (base at approximately 12.5 cm depth in core) are calculated as approximately 40,000 and 125,000 years B.P., respectively, possibly correlating with the global isotopic stages 5 and 4, and the lower part of stage 3. The maximum depletion in $\delta^{18}O$ occurs at about 60,000 years B.P., in isotopic stage 3.

The large-scale decreases in planktonic foraminifera and associated large depletions in $\delta^{18}O$ can be interpreted as the result of a major dilution of Arctic Ocean surface waters, where the surface salinity values decreased below the tolerance level of planktonic foraminifera. The presence of high pollen and spore concentrations suggests that this interval probably represents increased freshwater influx from major rivers draining lower latitudes during warmer intervals. River discharge probably provided the necessary organic compounds for blooms of gonyaulacoid dinocysts [Prakash, 1967]. The presence of Subarctic dinoflagellate species O. centrocarpum, S. ramosus, and Tectatodinium pellitum generally corresponds to light isotopic intervals and may indicate periods of increased advection of Atlantic and Pacific waters through the Fram and Bering straits during high sea-level stands.

Between 23 and 31 cm depth in core 103, low total planktonic foraminiferal counts correlate with very heavy oxygen isotopic values (Figure 7). This interval also contains very low pollen and spore and dinoflagellate concentrations. The ages of the upper and lower boundaries of this heavy $\delta^{18}O$ interval are estimated as 230,000 and 310,000 years B.P., suggesting a possible correlation of this interval with the global isotopic stage 8 (approximately 297,000 to 251,000 years B.P. [Shackleton and Opdyke, 1976]). The low abundances of all microfossils and the associated very heavy oxygen isotopic values suggest the formation of thicker and more extensive sea ice in the Arctic Ocean which, in turn, would decrease biological productivity.

Except for the uppermost few centimeters, the entire lithofacies L [Clark et al., 1980] in the CESAR cores

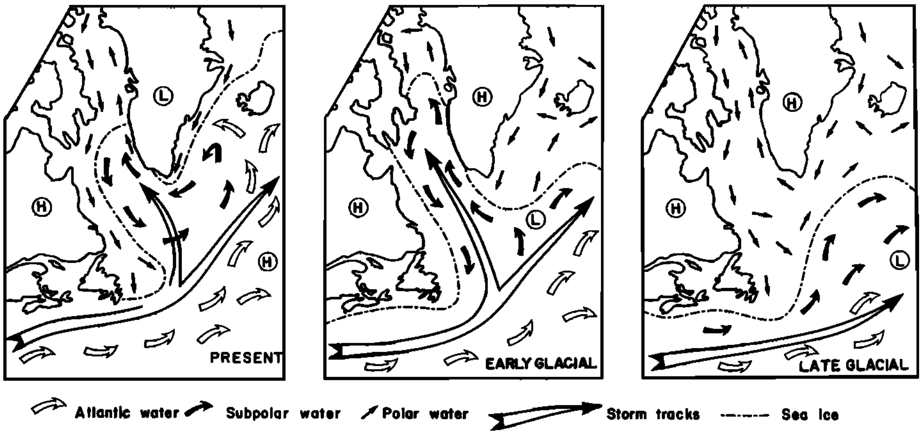


Fig. 8. Summary diagram showing major storm tracks, cyclonic and anticyclonic pressure cells (from Gates, 1976), inferred surface water circulation patterns and year-round sea ice cover. The following period are shown: (left) present, indicating interglacial periods, (middle) early glacial periods, and (right) late glacial periods.

is barren of foraminifera but includes considerable concentrations of tree pollen, spores, and dinoflagellates (Figure 7). The upper part of this interval correlates with very light $\delta^{18}O$ values. Age estimates using a constant rate of deposition suggest that the isotopically light interval between 38 and 32 cm depth in core 103 can be tentatively correlated with global isotopic stage 9. High abundances of pollen and spores and very light $\delta^{18}O$ values in this interval suggest significant freshwater influx to the Arctic Ocean. There is a progressive enrichment in $\delta^{18}O$ values from about 38 cm to about 30 cm depth in the core, which is associated with a progressive decline in pollen and spore concentrations and with a short interval of increased planktonic foraminiferal abundances. This trend may reflect gradual cooling of the global climate during the transition from isotopic stage 9 to stage 8, leading to extensive sea ice formation at the maxima of glacial stage 8 (Figure 7). Relatively high numbers of Quaternary palynomorphs in lithofacies L appear to reflect increased input of terrestrial organic matter from rivers flowing into the Arctic Ocean and relatively less sea ice, with higher primary production in the Arctic Ocean. The virtual absence of planktonic foraminifera in these sediments is difficult to explain, but it may reflect both low-salinity surface water and decreased carbonate preservation due to higher organic input from rivers.

WESTERN NORTH ATLANTIC PALEOCEANOGRAPHIC MODEL

The data from Baffin Bay [Aksu, 1981; Mudie and Aksu, 1984] and the Labrador Sea [Fillon and Aksu, 1985; Aksu and Mudie, 1985a; Aksu et al., 1986; de Vernal, 1986] can be used to reconstruct the paleoceanographic and paleoclimatic conditions that were associated with the large-scale continental ice sheet development in the seaway linking Labrador Sea and Baffin Bay (Figure 8). The correlations from south to north of peaks in Subarctic planktonic foraminifera and dinoflagellates and of faunal and floral abundances indicate that the surface waters over most of the Labrador Sea and southern Baffin Bay were not covered by perennial sea ice during the early to middle parts of major continental ice growth phases (Figure 8, middle). The influx of subarctic biota also implies significant advection of relatively warm surface waters concurrent with phases of the inception of glacial stages and preceding global glacial maxima. The duration of oceanic warmth during glacial stages appears to vary for different interglacial to glacial transitions. For example, data from isotopic stages 6 and 8 of several cores show little evidence for major oceanic cooling during the early and middle parts of these stages, and a corridor of oceanic warmth probably extended from the central North Atlantic [Ruddiman and McIntyre, 1981] into the northern Labrador Sea [Aksu and Mudie, 1985a] and southern Baffin Bay [Mudie and Aksu, 1984]. This oceanic configuration, with an intense thermal gradient between the ice-covered land and warmer ocean, creates an optimal condition for ice sheet growth by providing a local moisture source and by channelizing moisture-bearing lower-latitude storms into the Labrador Sea (Figure 8, middle) and supports the basic conditions of ice sheet growth postulated by Andrews et al. [1972], Johnson and McClure [1976], and Ruddiman et al. [1980].

Persistent tree pollen associated with relatively high pollen and spore concentrations in lower isotopic stages 6 and to a lesser extent 2, 4, and 8 suggest influxes of southerly air masses during early glacial periods (Figure 8, middle). The general decrease of pollen and spore concentrations and the disappearance of Alnus and Picea toward the later parts of the glacial stages correlate with the overall decline of Subarctic planktonic foraminifera and dinoflagellates, total radiolarians, coccoliths, and diatoms and may indicate a southerly shift of the Arctic-Atlantic air front and the subsequent predominance of the colder, dryer westerly Arctic air masses. A corresponding southerly shift of the oceanic Polar Front probably promoted extensive year-round sea ice formation in the Labrador Sea during the late glacial phases (Figure 8, right).

The transitions from glacial to interglacial periods in the cores studied are often marked by sediments barren of coccoliths, a decline in total planktonic foraminifera and radiolarians, and a sharp increase in diatoms. The decrease/disappearance of the fauna and flora at these intervals cannot be explained by calcium carbonate dissolution alone [Aksu and Mudie, 1985a], nor can it be explained by an overall decline in total biological productivity, because diatoms suggest relatively high productivity. The observed variations are best explained by the development of a low-salinity surface layer maintained by a continuous meltwater supply from decaying ice sheets and a renewal of Arctic Ocean inflow through the Arctic channels for a considerable amount of time.

CONCLUSIONS

The total data set for the transect extending from southern Labrador Sea to Baffin Bay thus confirms the ocean-atmosphere circulation model depicted for the Davis Strait foraminiferal and palynological data by Mudie and Aksu [1984] and Aksu and Mudie [1985a]. This model predicts that the onset of glaciations in and near eastern Canada is marked by a deepening of the northward path of the cyclonic storms, resulting in increased northward inflow of Subarctic waters to at least the Davis Strait and providing summer open water as a moisture source for the rapid buildup of the Laurentide and Innuitian ice sheets. This model is similar to that of Ruddiman and McIntyre [1981] who proposed that steepening of the land-sea thermal gradient at the inception of glacial stages would intensify the development of cyclonic storm tracks in the northwest Atlantic.

Several major features of the Arctic Ocean upper Quaternary record are tentatively correlated with the well-studied lower-latitude oceanic records. The reliable chronostratigraphic framework that is now established for the CESAR cores, however, lead to several questions that must be answered before a definite model can be estabished for the late Cenozoic history of the Arctic Ocean. The main questions are listed as follows: (1) Why can a large meltwater influx (either fluvial or glacial) be discerned for interstadial stage 3, but not for transitions from glacial to interglacial stages 7/8 and 5/6 or for early interglacial periods? (2) Why does the Arctic Ocean record only indicate very extensive sea ice formation (i.e., very heavy oxygen isotopic values) in glacial isotopic stage 8? (3) Does the restricted stratigraphic distribution of pteropods indicate relatively recent migration into the Arctic Ocean, or does it reflect a late Pleistocene change in carbonate preservations and/or an increase in the depth of the aragonite lysocline? (4) What is the reason for the virtual absence of planktonic foraminifera in the early Pleistocene, despite the presence of calcareous benthic foraminifera and calcareous dinocysts? (5) What is the paleoecological significance of the dramatic faunal turnover from exclusively arenaceous to predominantly calcareous benthic foraminifera in the early Pleistocene?

The answers to these questions are crucial to a full understanding of the role of the Arctic Ocean in northern hemisphere upper Cenozoic climatic evolution and for development of globabl climatic models that include ocean-atmosphere-cryosphere feedback and teleconnections between the Arctic and high-latitude North Atlantic ocean basins.

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A. E. Aksu and S. A. Macko, Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland, Canada A1B 3X5.

A. de Vernal, Département des Sciences de la Terre, Université du Québec à Montréal, Casse Postale 8888, Succursale A, Montréal, Québec, Canada H3C 3P8.

P. J. Mudie, Atlantic Geoscience Centre, Geological Survey of Canada, Bedford Institute of Oceanography, P. O. Box 1006, Dartmouth, Nova Scotia, Canada B2Y 4A2.

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