8 May 1981, Volume 212, Number 4495

SCIENCE

Oceanic Mechanisms for Amplification of the 23,000-Year Ice-Volume Cycle

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Summary. Situated adjacent to the largest Northern Hemisphere ice sheets of the ice ages, the mid-latitude North Atlantic Ocean has an important role in the earth's climate history. It provides a significant local source of moisture for the atmosphere and adjacent continents, forms a corridor that guides moisture-bearing storms northward from low latitudes, and at times makes direct contact along its shorelines with continental ice masses. Evidence of major ice-ocean-air interactions involving the North Atlantic during the last 250,000 years is summarized. Outflow of icebergs and meltwater initially driven by summer insolation over the ice sheets affects mid-latitude ocean temperatures, summer heat storage, winter sea-ice extent, and global sea level. These oceanic responses in turn influence the winter moisture flux back to the ice sheets, as well as ablation of land ice by calving. Spectral data indicate that the oceanic moisture and sea-level feedbacks, in part controlled by glacial melt products, amplify Milankovitch (insolation) forcing of the volumetrically dominant mid-latitude ice sheets at the 23,000-year precessional cycle.

Milankovitch (*1*) hypothesized that ice ages begin when summer insolation at critical latitudes in the Northern Hemisphere (~ 65°N) decreases to values low enough to cause the snow line to descend and intersect large regions of high terrain. Under such conditions, large fractions of snow and ice survive from the preceding cold season. As the ice cover grows, increased reflectivity aids the glaciation process.

The orbital configuration that Milankovitch considered critical for rapid growth of the ice sheets in the Northern Hemisphere is shown in Fig. 1A. Summer insolation in the Northern Hemisphere is at a minimum because (i) summer occurs at the most distant pass (aphelion) in the earth's slightly eccentric orbit around the sun, and (ii) the tilt is small, which decreases summer insolation in high northern latitudes. North of 65°N, the tilt factor is more important in minimizing summer insolation; south of 65°N, the precessional position at aphelion is dominant.

Three decades of research on deep-ocean sediments have established a record of ice-volume change that supports the Milankovitch hypothesis. The ratio of two stable isotopes of oxygen (^{18}O and ^{16}O) in shells of benthic foraminifera primarily reflects the growth and decay of continental ice (*2*). Isotopic records covering the past 250,000 years (Fig. 2) are based on benthic foraminifera, which minimize temperature variations and render the best available records of ice volume (*3*).

Four major transitions toward heavier isotopic values (larger ice volume) are evident in cores from both the Pacific and Atlantic oceans (Fig. 2). The following four transitions have the largest amplitude and speed of change: the boundary between stages 5 and 4 dated at 75,000 to 72,000 years before present (B.P.); the boundary between substages 5e and 5d dated at 115,000 years B.P.; the boundary between stages 7 and 6 dated at 190,000 to 185,000 years B.P.; and the boundary between substages 7c and 7b dated at 230,000 years B.P. (*4*).

Northern Hemisphere summer insolation at the top of the atmosphere at the latitudes of the major ice sheets (Fig. 2) varies with dominant frequencies of 41,000 years north of about 65°N and 23,000 years south of 65°N. The four largest ice-growth transitions correlate with the strongest minima in summer insolation at 65°N (*1, 5*). Those lesser ice-growth phases not directly explained by the summer insolation curve at 65°N seem to be accounted for by radiation minima farther south at latitudes 45°N to 60°N. This correlation, however, does not prove the mechanism proposed by Milankovitch.

Hays *et al.* (*6, 7*) demonstrated that known frequencies in the earth's orbital variations are recorded in ice-volume and oceanic responses at the earth's surface. Stressing only the statistical association, they avoided the problem of selecting specific mechanisms by which orbitally controlled insolation changes alter climate. Imbrie and Imbrie (*8*) demonstrated that the summer insolation curves at 65°N and 45°N can reproduce isotopic ice-volume records, with allowance for what Milankovitch called the retardation effect (slow ice-sheet response times for growth and decay). They cautioned that even these statistical correlations do not define a physical mechanism by which climatic changes occur.

Milankovitch (*1*) focused on insolation and considered the energy flux caused by atmospheric and oceanic circulations as secondary perturbations. However, this flux carries the energy and moisture necessary for ice-sheet growth. Most of the world's heat is stored in the oceans, and a significant portion of the transport

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0036-8075/81/0508-0617$02.00/0

across latitudinal belts is oceanic (*9*). We describe evidence that changes in the amount of moisture made available to the ice sheets from the North Atlantic Ocean form a second major mechanism in the growth and decay of Northern Hemisphere glaciers. We also cite evidence that changes in sea level affect ice-sheet size through the mechanism of iceberg calving.

Ice Growth: Oceanic Evidence

Our research (*10*) has defined two conditions that accompany rapid and large-scale growth of Northern Hemisphere ice sheets.

1) Ice-rafted deposition in the North Atlantic subpolar gyre is low during the early and middle stages of ice growth and may decrease from higher values reached during periods immediately preceding ice growth. Only when ice growth slows or ceases does ice rafting reach high values. The low rate of deposition of ice-rafted detritus during glacial growth is consistent with the Milankovitch hypothesis. It implies that more ice was retained on the continents during glacial growth, and ice retention on the continents as a means of promoting a positive glacial mass balance is the centerpiece of the Milankovitch hypothesis.

2) Surface waters over large parts of the subpolar and the northern subtropical Atlantic remained warm and saline for several thousand years into the major ice-growth phases (*10, 11*). Between 50°N to 65°N, this warm-ocean lag is in the range of 1000 to 5000 years, and the region of persisting warmth appears to vary for different ice-growth transitions. Figure 3 shows the oceanic configuration at the isotopic 5-4 ice-growth transition, 75,000 to 72,000 years B.P., with detectable oceanic warmth in all cores south of Iceland (*10*). At the isotopic 5e-5d ice-growth transition, some 115,000 years B.P., a corridor of greatest warmth extended northward along the eastern side of the Atlantic into the Norwegian Sea at 68°N (*10, 12*).

The northwestern part of the subtropical gyre (Fig. 3) maintains or even increases its warmth and salinity throughout much of the interval of rapid ice growth (*10, 11*). This oceanic warmth lags behind ice growth by 5000 years or more; it recurs on the four major ice-growth phases studied.

Role of the North Atlantic Ocean in Ice Growth

Lingering warmth and high salinity in the high-latitude subpolar and northern subtropical gyre are not encompassed by the Milankovitch hypothesis. Incident radiation in the summer half-year is at a minimum, as is the integrated annual radiation (*5*). In addition, the growing ice sheets cool the high latitudes by increasing Northern Hemisphere albedo, and the atmosphere extracts energy and moisture from the ocean for rapid glacier growth.

This lingering oceanic warmth creates an optimal configuration for rapid ice-sheet growth by providing moisture from both local and low-latitude sources (*10*). Moisture will be extracted locally from the subpolar and northern subtropical Atlantic because of its high heat content. In addition, the high salinity and high winter insolation (Fig. 2) keep the ocean relatively free of sea ice, creating a large surface of open ocean from which to extract moisture in winter.

On a more regional scale, this configuration (Fig. 3) provides an oceanic corridor for storms. The intense thermal gradient between the ice-covered land and the warm ocean promotes a storm track which directs low-latitude storms northward. At the stage 5-4 ice-growth transition (Fig. 3), we infer that the strong thermal gradient off Newfoundland guided storms northward into the Labrador Sea toward the heart of the growing Laurentide Ice Sheet (*13*).

Whether of local or low-latitude origin, precipitation on high northern latitude continents was more likely to fall as snow than as rain because of reduced summer insolation (*1*). Winter was the main time for moisture to be extracted from the ocean and precipitated on land, and high winter insolation could enhance this process (*14*).

Ice Decay: Oceanic Evidence

Examination of glacial terminations II and I (the boundaries of isotopic stages 6-5 and 2-1) has yielded evidence that the subpolar North Atlantic, at least as far south as 50°N, was virtually barren of coccoliths, planktic foraminifera, and diatoms for several thousand years (*15*). The concentrations of these microfossils drop by more than one order of magnitude across the early parts of these deglacial transitions. With reasonable allowance for sediment mixing, the decrease probably amounts to several orders of magnitude and may indicate a gyre whose surface waters were barren of planktic life for several thousand years.

The surface-water regime that best explains low productivity throughout the year is one with a low-salinity meltwater layer maintained by yearly additions of large volumes of low-nutrient fresh water from the melting of surrounding continental glaciers. The North Atlantic is the major receptacle of Northern Hemisphere glacial meltwater, and isotopic data (Fig. 2) mark terminations II and I

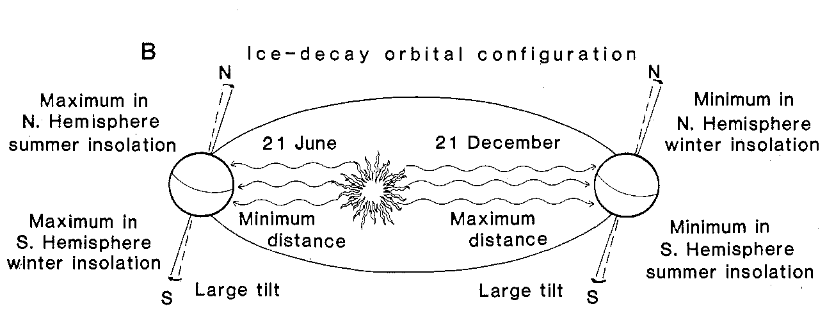

Fig. 1. (A) Orbital configuration of the earth and sun during periods of rapid ice growth based on Milankovitch (*1*). Critical summer season of minimal insolation and ice melting in the Northern Hemisphere is related to low tilt and 21 June aphelion precession position (distant pass in the slightly eccentric orbit). (B) Orbital configuration during ice disintegration, as suggested by Emiliani and Geiss [in (*22*)] and Broecker (*41*). Maximum tilt and perihelion (close-pass) 21 June precessional position favor high summer insolation and rapid ice disintegration in the Northern Hemisphere.

as strong deglaciations. Meltwater during high-insolation summers must have inhibited productivity by maintaining a stably stratified low-salinity surface layer (16) with a strong basal pycnocline that prevented nutrient-rich waters at depth from mixing with the sunlit near-surface waters.

In addition, sand-sized ice-rafted detritus reaches very high percentages in the zones barren of microfossils, but sedimentation rates remain nearly constant (15). This indicates that the deficit of planktic shells from surface waters is largely offset by increased delivery of ice-rafted debris.

South of the microfossil-barren subpolar ocean, the sea surface reached minimum temperatures in northern subtropical latitudes at 35°N to 45°N (Fig. 4). These minima correlate with the early to middle stages of ice-sheet disintegration on land (for example, 16,000 to 13,000 years B.P. in the last deglaciation) (15).

Role of the Ocean in Ice Decay

The earth's orbital configuration during strong deglaciations (terminations II and I) (Fig. 1B) and astronomical data (Fig. 2) indicate a summer insolation maximum of unusual strength over all ice-covered latitudes of the Northern Hemisphere. Broecker *et al.* (17) described the close correlation and implied causal connection between the last two abrupt deglaciations evident in isotopic records and the strong summer insolation maxima (18). This relation suggests rapid ice disintegration under the strong summer sun, with large volumes of meltwater and icebergs flowing to the ocean.

The intense summer insolation maxima at high northern latitudes during deglaciations are roughly balanced by comparably intense winter insolation minima (Fig. 2). Because unusually low winter insolation during terminations II and I coincides with a maximum influx of low-salinity meltwater, we suggest that sea ice covered a very large area of the subpolar North Atlantic (southward to at least 50°N) during deglacial winters (15, 19). The winter sea-ice cover plays a supplementary role in explaining the low productivity during terminations II and I. Without a sea-ice cover, the cooling during winter by cold dry masses of polar air would cause deep convection, bringing nutrient-rich waters to the surface. Winter icebergs probably helped minimize wind mixing and enhance sea-ice formation by calming the sea surface.

Icebergs also influence the cooling of the ocean surface by extracting energy both for the latent heat of melting as well as that required for warming. We have evidence that iceberg calving was the major mode of rapid Northern Hemisphere deglaciation from 16,000 to 13,000 years B.P. (15). This calving process is thought to be activated initially by summer melting of the major mid-lati-

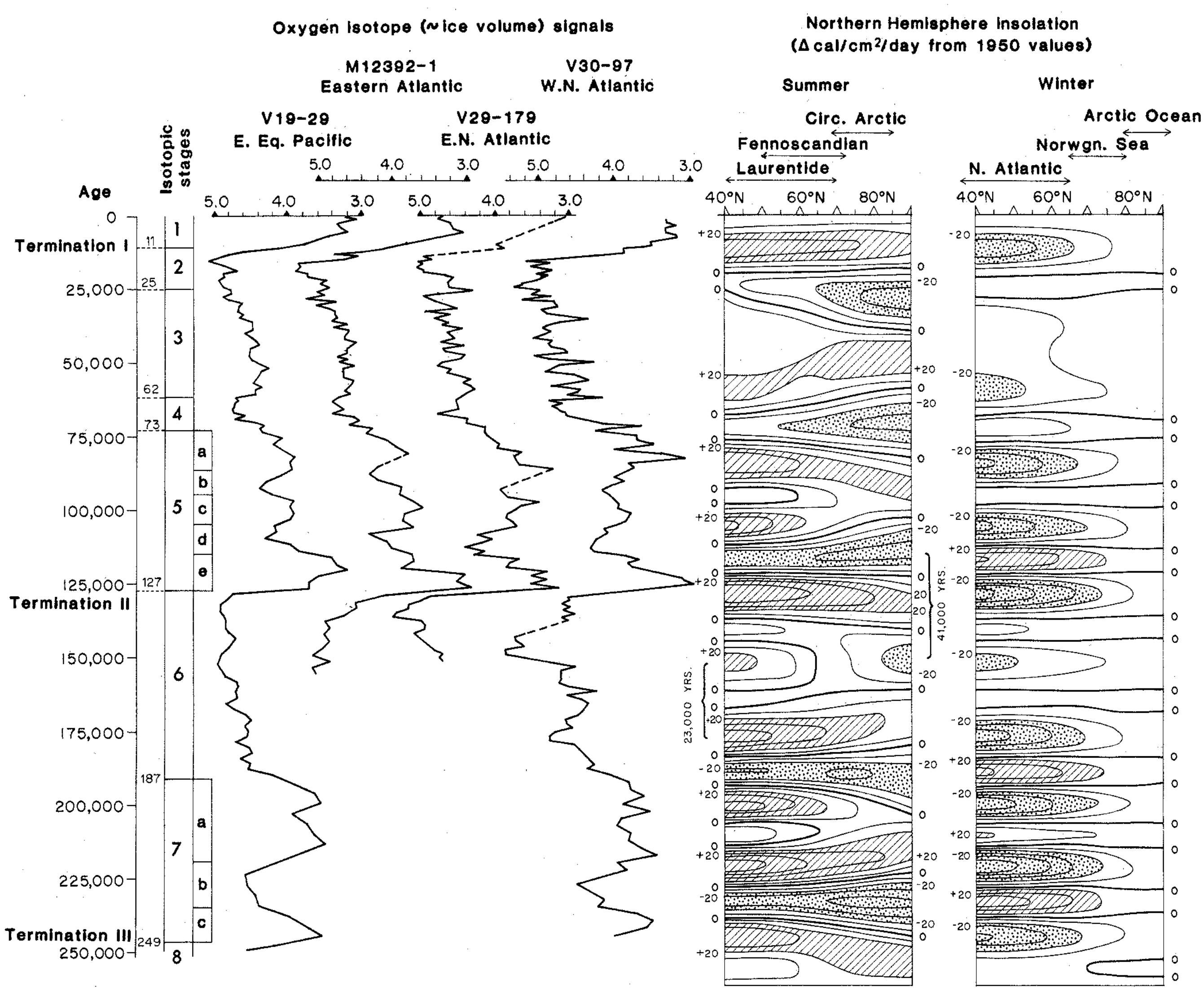

Fig. 2. Comparison of isotopic (~ ice volume) curves and astronomical calculations of summer and winter insolation (1, 5). Ice-volume curves are based on oxygen isotopic records measured on benthic foraminifera in cores from widely separated regions (3). Ice-growth phases correspond to low summer insolation and high winter insolation; ice-decay phases correspond to high summer insolation and low winter insolation. Latitudinal extents of ice sheets and ocean basins shown above summer and winter insolation plots, respectively.

tude ice sheets under rising summer insolation (*20*). Then, through a positive feedback loop, rising sea level causes increased iceberg calving which in turn hastens sea-level rise (*20*).

The major impact of this iceberg influx is on the North Atlantic from 40°N to 55°N, where maximum abundances of ice-rafted debris are deposited (*21*). We estimated that the melting of icebergs in this area could chill the upper 100 meters of the ocean by several degrees during early stages of rapid deglaciations (*15*). Thus, iceberg melting appears to be significant in the mid-latitude oceanic temperature minimum on deglaciations (*22*). Oceanic cooling could also be enhanced by cold dry winds blowing from the high-albedo surface of the still-extensive Laurentide Ice Sheet (*23*). The icebergs also provide fresh water for the low-salinity meltwater layer.

The observed relationships between summer and winter insolation, ice volume, winter sea-ice formation, and ocean temperature during periods of ice decay (terminations II and I) are summarized in Fig. 4. Both the meltwater and iceberg influxes influence the moisture flux around the mid-latitude North Atlantic.

Adam (*16*) suggested that density stratification caused by glacial meltwater will confine the storage of summer heat to shallow, and therefore volumetrically insignificant, layers of the mid-latitude North Atlantic Ocean. Studies of recent annual changes in the Labrador Sea support this concept (*24*). The iceberg-induced chilling of ocean-surface waters during deglaciation will diminish still further the summer storage of heat in the North Atlantic Ocean. These two factors will suppress the transport of moisture in winter from local oceanic sources to the ice sheets (*16*).

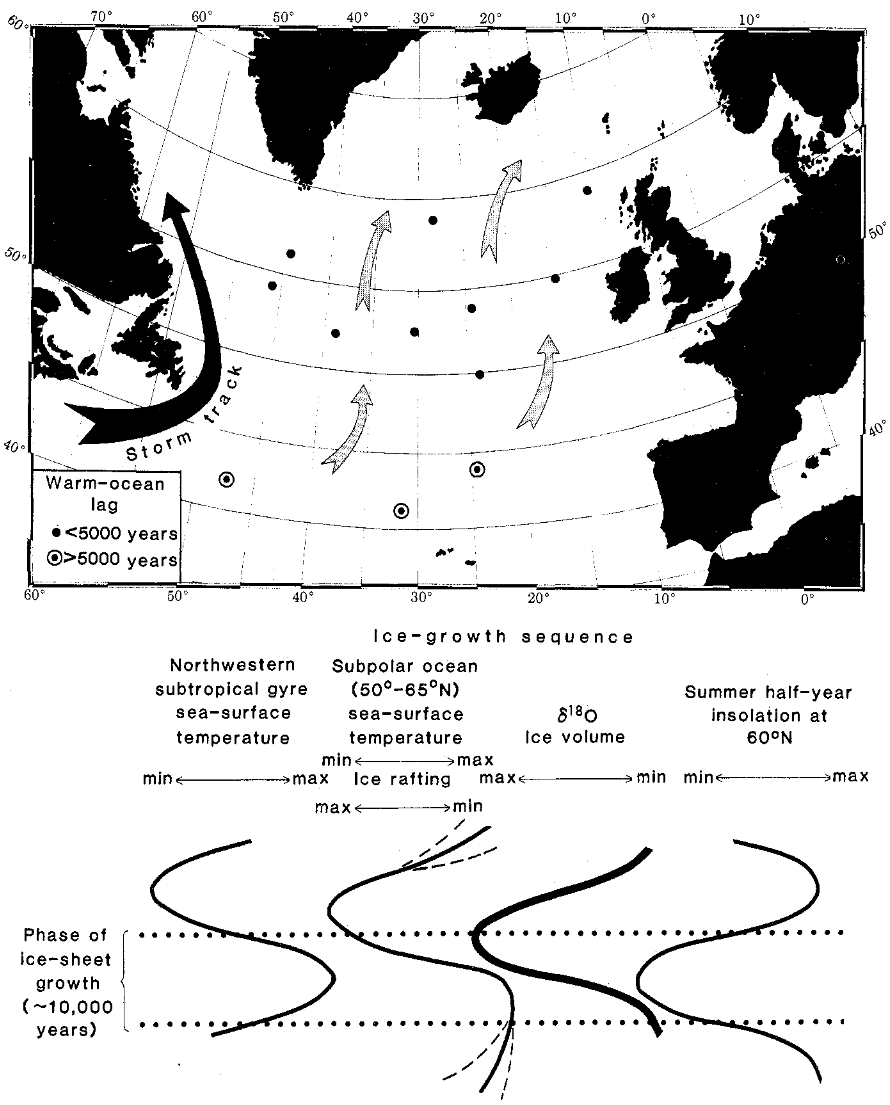

Fig. 3. Sequence of North Atlantic changes associated with major ice-growth phases. Map shows location of cores (circles) indicating oceanic regions with lagging warmth and high salinity and with ice-rafting minima during glacial buildup (*10, 11*). Arrows indicate northward moisture transport in winter over a warm ocean free of sea ice.

The existence of an extensive region of winter sea ice across the subpolar North Atlantic during deglaciation also implies a reduced supply of moisture to the Northern Hemisphere ice sheets. This reduction applies both to moisture extracted locally from mid-latitude waters as well as to that brought in by storms from low latitudes.

Local moisture loss is implicit in a sea-ice lid extending southward to at least 50°N, because air masses extract the greatest amount of moisture from the mid-latitude ocean in winter. Moisture-bearing storms will continue to form over the lower latitude ocean during ice-decay intervals; however, the winter sea-ice limit marks such a strong temperature boundary that its effects on atmospheric circulation will oppose northward movement of these storms. Because the cyclonic storm track tends to follow the strongest albedo boundaries and thermal gradients (*25*), the winter storms should move eastward along the strong thermal boundary at the edge of the sea-ice limit shown in Fig. 4. On balance, this will keep the storms away from all the Northern Hemisphere ice sheets during early and middle phases of deglaciation. Those that do move northward will be cut off from their source of energy and will weaken.

We summarize in Fig. 5 the major feedback mechanisms by which we believe the ocean enhances the rate of ice disintegration during rapid deglaciations. These loops link the interactions of insolation, land ice, the ocean, and the atmosphere at mid-latitudes of the Northern Hemisphere.

Spectral Analysis of North Atlantic Climatic Response

Evidence from intervals of rapid ice growth or decay omits large segments of the climatic record. To extend our investigation of the North Atlantic record, we analyzed one core in detail through the last 250,000 years: core V30-97 (41°00′N, 32°55.5′W, 3371 meters), which underlies the northern part of the subtropical gyre just beyond the southern limit reached by polar water during maximum glaciations (Fig. 4).

Core V30-97 (Fig. 6) yielded a high-resolution record of changes in oxygen isotopes and in sea-surface temperatures from examination of benthic and planktic foraminifera, respectively (*26*). The isotopic record compares favorably with records from the lower latitude Atlantic and equatorial Pacific (Fig. 2). Because benthic foraminifera needed for isotopic analysis are rare in the North Atlantic during the last 15,000 years, we used a wide-diameter box core (V30-101K, 44°06′N, 32°30′W, 3519 m), located 300 kilometers north of core V30-97, to obtain a record across this interval. The two curves were spliced together from linear interpolations of carbon-14 dates; the isotopic values and species analyzed at the spliced levels match well (Fig. 6). Estimates of sea-surface temperatures and values of calcium carbonate are from core V30-97.

Using the chronology of Kominz and Pisias (*7*) to align the isotopic stage boundaries, we ran a spectral analysis of both the isotopic and sea-surface temperature data (*27*). The isotopic record yielded a strong spectral peak around 100,000 years, with weaker peaks around 41,000 and 23,000 years (Fig. 7); all are orbital frequencies detected in deep-sea cores (*6, 7*). In this respect, our results merely corroborate the earlier finding that ice volume in part fluctuates with orbital periodicities of 100,000, 41,000, and 23,000 years.

Spectral analysis of sea-surface temperatures shows a strong 100,000-year peak which must be considered ill-constrained because the record incorporates only two-and-one-half 100,000-year cycles (Fig. 7). The 23,000-year peak is, however, highly significant, incorporating more than ten 23,000-year cycles. The 23,000-year peak far exceeds the 95 percent confidence level (Fig. 7); spectral power this intense has not been reported from deep-sea cores at any frequency (*28*). The amount of total variance under the 23,000-year peak is 45 percent.

Core V30-97 is from the northern subtropical gyre, the area indicating lingering warmth during ice growth and lingering cold during ice decay. This spectral evidence shows that sea-surface temperatures in the northern subtropical gyre not only lag the ice-sheet response but also fluctuate to a specific beat—the orbital precession cycle of 23,000 years.

Figure 8 shows the phase relationships of three parameters: (i) the precessional index cycle referenced by convention to be in phase with summer insolation values at 21 June in the Northern Hemisphere; (ii) the filtered 23,000-year component of the summer sea-surface temperature curve in core V30-97 (Fig. 6); and (iii) the filtered 23,000-year component of the ice volume ($\delta^{18}O$) curve in core V30-97 (*29*).

We derived mean values for the phase relationships (*29*) across the available record (~ 185,000 years after truncation

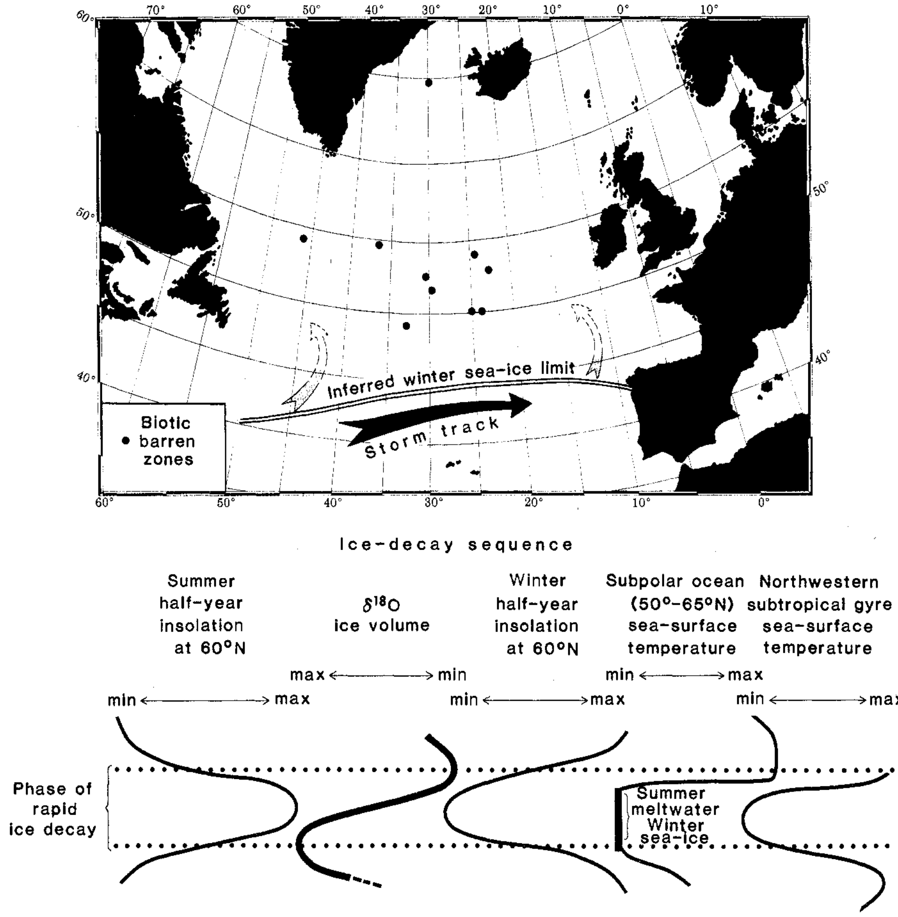

Fig. 4. Sequence of North Atlantic changes associated with major ice-decay phases (terminations). Map shows location of cores (circles) with biotic barren zones indicative of summer meltwater layer and winter sea ice during deglaciation (*15*). Arrow indicates winter storm track eastward along pack-ice limit.

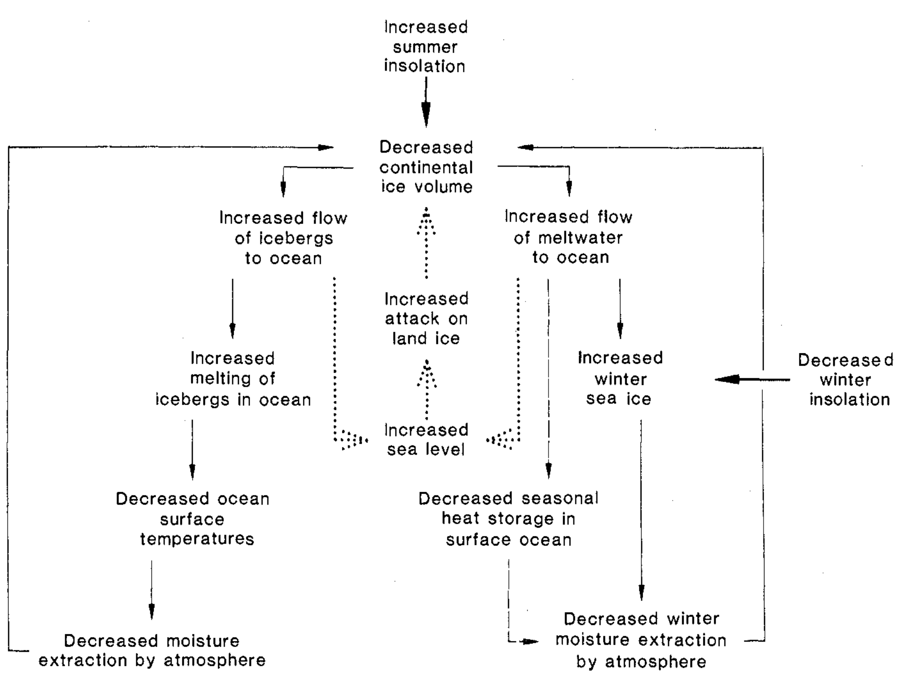

Fig. 5. Oceanic feedback loops that hasten ice disintegration are driven initially by high summer insolation. Sea-level mechanism (dotted lines) is discussed in (*20*); heat-storage loop (dashed lines) was proposed by Adam (*16*). Melting of icebergs (loop on left side) also creates meltwater, enhancing the effectiveness of the other two moisture feedback loops. During ice-growth phases, these feedback loops permit moisture to reach the ice sheets and diminish or stop the sea-level attack. These processes amplify Milankovitch (summer insolation) forcing at the 23,000-year precessional cycle.

at each end): (i) Northern Hemisphere summer (21 June) insolation at the precessional frequency leads the 23,000-year component of ice volume by 3000 ± 1500 years; (ii) the 23,000-year component of ice volume leads the 23,000-year component of sea-surface temperature by 6000 ± 1500 years; and (iii) summer (21 June) insolation leads the 23,000-year component of sea-surface temperature by 9000 ± 1500 years.

These phase relationships confirm the relationships of insolation, ice volume, and sea-surface temperature depicted in Figs. 3 and 4 but attach these relationships to the 23,000-year precessional cycle. The lag of sea-surface temperature behind ice volume (6000 years) is about one-quarter of a 23,000-year wavelength; thus, the phases of ice-sheet growth ($\delta^{18}O$ decrease) are marked by low sea-surface temperatures (Fig. 8). The 9000-year lag of sea-surface temperature behind summer insolation represents almost a half-cycle phase lag for the 23,000-year period. The record of sea-surface temperature at this frequency is almost in phase with winter solstice insolation (21 December), as shown in Fig. 4; it falls directly in phase with November insolation.

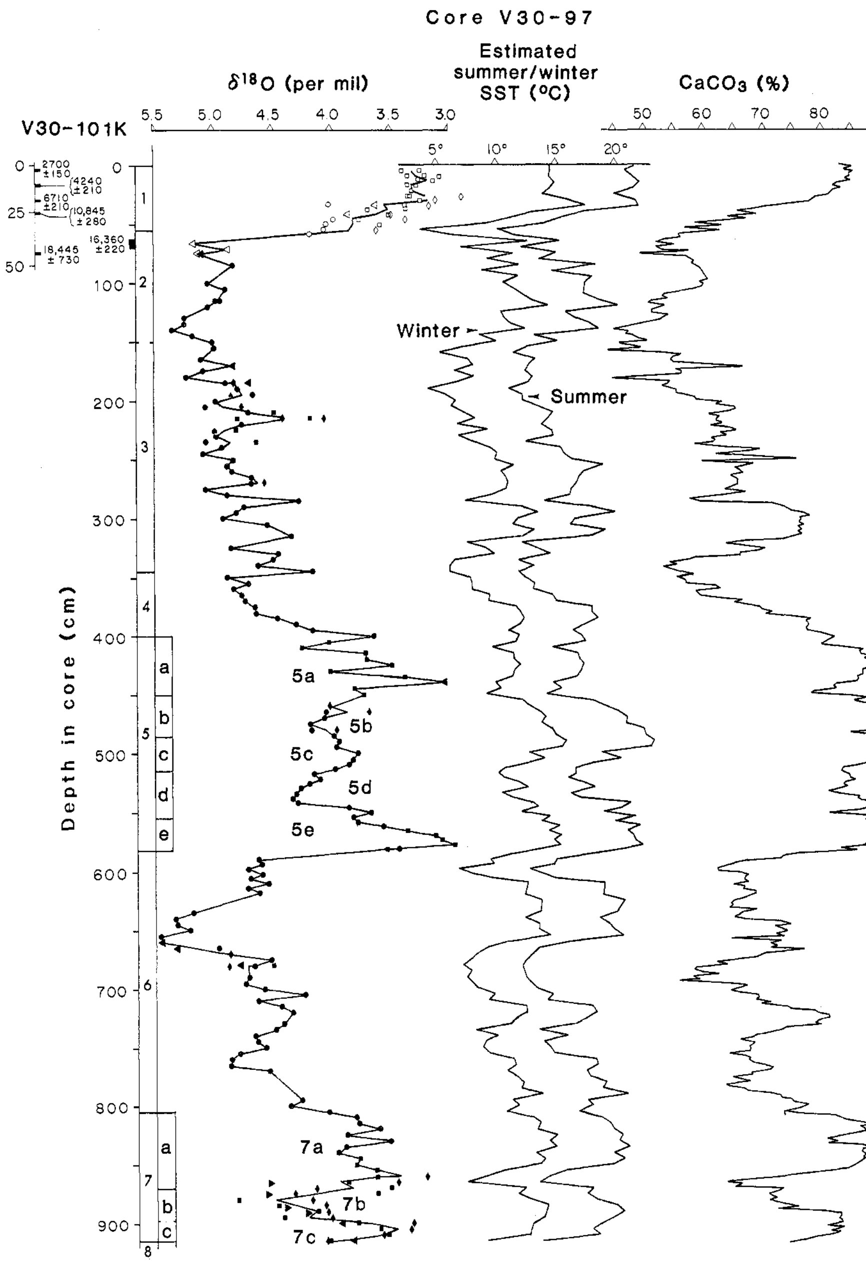

Fig. 6. Downcore record of $\delta^{18}O$, the per mil enrichment of ^{18}O, (~ ice volume), estimated summer and winter sea-surface temperature, and $CaCO_3$ in northern subtropical gyre core V30-97 (*26*). Isotopic values are based on *Uvigerina peregrina* (circles) and three species referenced to PDB and corrected to *Uvigerina*: squares are *Cibicides wuellerstorfi* (+0.64), diamonds are *Pyrgo* spp. (−0.51), and triangles are *Melonis* spp. (+0.40). Values from core V30-101 K (open symbols) were spliced with values from core V30-97 (filled symbols) by ^{14}C control. Temperature estimates are based on transfer function derived by Ruddiman and Glover [in (*23*)]. The $CaCO_3$ analyses are precise to ±0.5 percent. Isotopic stages 1 to 8 are indicated in the column on the left.

Ice Melt in the North Atlantic

The dominant 23,000-year frequency in the record of sea-surface temperatures (Figs. 6 to 8) must be caused by orbital changes in insolation. To invoke insolation as a cause of these temperature changes, it is necessary to assume that the orbitally controlled radiation changes received by the ocean in one hemisphere or season is somehow able to overwhelm the influence of the opposing radiation differential in the opposite hemisphere or season. To overcome the tendency toward seasonal and hemispheric counterbalancing, specific mechanisms that amplify or redistribute the observed insolation variations are required.

There are two complementary ways to address this problem. One would be to identify a low-latitude mechanism that periodically forces the northward delivery of oceanic energy or heat; the other would be to identify a high-latitude mechanism that periodically obstructs the amount of oceanic energy or heat that moves into higher northern latitudes. At this point, we lack the requisite low-latitude oceanic evidence to evaluate the first possibility and will only explore the second.

The strength of the 23,000-year spectral peak and lack of a 41,000-year peak in core V30-97 eliminates any mechanism that depends on initial orbital forcing acting poleward of roughly 65°S or 65°N because of the dominance of the 41,000-year obliquity (tilt) cycle at these high polar latitudes. It favors forcing tied to the precession cycle and initially affecting the earth somewhere between 0° and 65°N or 0° and 65°S, the latitude ranges in which precession dominates over obliquity.

Our investigation of the North Atlantic during ice growth and decay (Figs. 3 and 4) has defined several Northern Hemisphere mechanisms that amplify insolation changes (Fig. 5). We now suggest that these mechanisms are feasible in spectral terms at the 23,000-year precession cycle.

The Laurentide Ice Sheet, the largest and most southerly of the Northern Hemisphere ice-age ice sheets (Fig. 2), at maximum size spans latitudes 40°N to 70°N. Although the ice-growth centers are not known, the ice-retreat margins of the last deglaciation indicate a Labrador-

Ungava remnant centered at 55°N and a Keewatin remnant at 65°N (*30*).

At all latitudes north to just beyond 65°, summer insolation at the top of the atmosphere is dominated by the 23,000-year precessional cycle (Fig. 2). Spectral analysis of 60°N summer insolation shows more 23,000-year than 41,000-year (tilt) power during the last 250,000 years (*6*). If we accept Milankovitch's hypothesis that summer insolation controls ice volume, this means that most of the Laurentide Ice Sheet is subject to a predominantly 23,000-year cycle of growth and decay. The southeastern extreme of this ice sheet, the eastern Laurentide region, is subject to the strongest and spectrally most pure 23,000-year insolation cycle of any ice sheet and is also the most proximal body of land ice draining into the ocean flow leading to core V30-97. Thus, the growth and decay cycle of the Laurentide ice mass is implicated in the tempo of sea-surface temperature response at core V30-97 (*31*).

We have noted the impact that meltwater and icebergs can have on the oceans during glacial terminations. The high insolation apparently produces a maximum influx of meltwater and icebergs over much of the mid-latitude Atlantic (35°N to 65°N). We suggest that a smaller but still significant version of the same effect will occur on every 23,000-year cycle of rising and maximum summer insolation.

The greatest impact of these summer insolation maxima will fall on the lowest latitude ice sheet, the eastern Laurentide (or Labrador-Ungava) sector. Disintegration of this ice mass should feed icebergs and meltwater to the North Atlantic along a roughly zonal (west-to-east) path at 50°N. Although the modern Labrador Current generally stays close to the coast of North America, the large volumes of meltwater involved during deglacial cycles would presumably have spread much farther east (Fig. 9). Ice-rafted debris gives evidence of significant eastward penetration of icebergs during much of the last 120,000 years (*21*). Using the argument applied to the terminations, we note that the 23,000-year maxima of summer insolation are matched by nearly comparable winter insolation minima (Fig. 2). Thus, the tendency to form winter sea ice in the western mid-latitude North Atlantic will also be dominated by a 23,000-year cycle.

The hypothesis of meltwater and iceberg control of estimated sea-surface temperatures in this part of the subpolar North Atlantic (*32*) can explain the one-quarter wavelength (6000-year) lag of the temperatures behind ice volume at the 23,000-year precessional frequency. The one-quarter wavelength phase lag specifies that sea-surface temperature varies in phase with the first derivative of ice volume, which is the meltwater and iceberg flux.

Donn and Ewing (*33*) interpreted the apparent lag in the oceanic transition out of the last glaciation as evidence that insolation is not the key to climatic change. This conclusion was based on the argument that the more responsive oceanic surface waters would not lag behind the less responsive ice sheets if insolation were the critical forcing function. Our evidence shows that insolation can explain the delayed oceanic response of the mid-latitude North Atlantic if the influences of the meltwater and iceberg flux are taken into account (*34*).

We suggest that the deglacial outflow formed a plume of low-salinity, iceberg-laden water emanating from the main

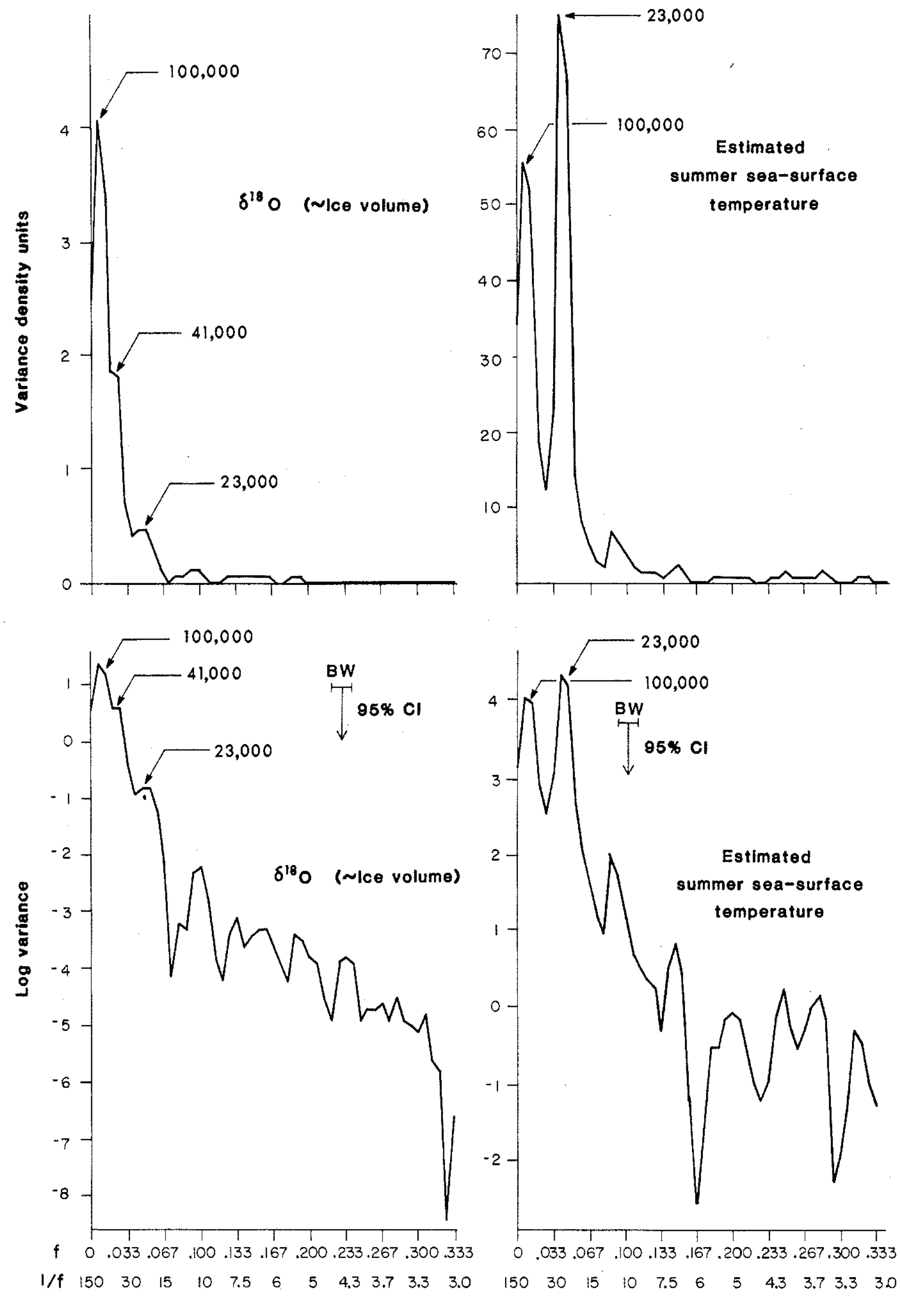

Fig. 7. Spectra of $\delta^{18}O$ and sea-surface temperatures in core V30-97 (*27*). Note the strong 23,000-year peak in the temperature record. Number of samples is 165; number of lags (*m*) is 50; sampling interval (Δt) is 1500 years. Plots on top show variance density units as a function of frequency. Log-linear plots (bottom) show the natural log of the variance as a function of frequency. A confidence interval (*CI*) at the 95 percent level, calculated from the chi-square distribution for $2n/m$ degrees of freedom is shown together with the bandwidth (*BW*) for a Hamming lag window ($1.258/m\Delta t$ cycles per 1000 years).

source area (Fig. 9). At times of minimal outflow (for example, during today's ice volume minima and during ice growth), the lack of summer icebergs and meltwater and winter sea-ice plumes enables the warm, saline Atlantic water to dominate the surface flow even in the Labrador and Norwegian seas. But with increasing melt-product outflow, the plume of icebergs and low-salinity water would spread farther to the south and east from the dominant Laurentide source (Fig. 9). This outflow might be joined by 23,000-year outflow from the southern Fennoscandian Ice Sheet after large glaciations (*31*).

The culmination of the meltwater and iceberg outflow comes at the high-amplitude precessional peaks centered on glacial terminations (Fig. 4). At this point, both the Laurentide and other outflows reach such intensities that they coalesce and send icebergs and meltwater south and east across the entire subpolar gyre to the Bay of Biscay (Figs. 4 and 9).

The potential southward and eastward extent of winter sea ice also fluctuates in large part with a 23,000-year cycle and follows more or less the meltwater-iceberg pattern shown in Fig. 9. Under the extremely low winter insolation characteristic of terminations, we infer that even the southeastern corner of the subpolar North Atlantic in the Bay of Biscay would freeze in winter.

Moisture Feedback to Land Ice

The North Atlantic from 40°N to 65°N represents only about 2 to 3 percent of the earth's surface, but most of the large Northern Hemisphere ice sheets are arrayed around its perimeter (Fig. 9). The subpolar to subtropical North Atlantic is thus the nearest major moisture source for glacier growth. The moisture flux available to the ice sheets is largely dependent on this ocean, particularly in the winter half-year of strongest moisture extraction.

The feedback loops presented in Fig. 5 suggest that the amount of warm ice-free water in contact with the atmosphere in the mid-latitude and high-latitude North Atlantic during winter is inversely related to the arriving summer melt-product flow. Similarly, the likelihood of low-latitude storms moving northward over the North Atlantic is inversely related to the deglacial outflow. This suggests that summer melt-product flux from the ice sheets in large part determines the winter oceanic moisture feedback to the ice sheets. Ice sheets thus starve themselves for winter moisture from the oceans when summer insolation is causing them to disintegrate most rapidly; conversely, a nearby ice-free source of abundant winter moisture is maintained in the ocean adjacent to the ice sheets when summer insolation is allowing glacial ice to accumulate most rapidly on the continents.

The moisture feedback loops (Fig. 5) apply to the region near core V30-97 specifically at the 23,000-year precessional frequency. At this frequency, this part of the North Atlantic acts as a perfectly phased moisture amplifier of Milankovitch's summer insolation mechanism, nourishing ice-sheet accumulation phases and starving ice-disintegration phases.

Although numerous glaciation hypotheses (*35*) involving oceanic moisture have touched on some of the elements in the moisture feedback loops (Fig. 5), two critical elements have been ignored: winter insolation and winter sea ice. Also, the moisture feedback loops have not been tied specifically to the 23,000-year precessional cycle.

It is not clear how much of the high-latitude North Atlantic participates in these moisture feedback loops at the 23,000-year periodicity. Presumably, the effect is strongest in plumes marking the axis of the Laurentide iceberg and meltwater outflow and dies away progressively to the south, east, and north (Fig. 9).

These schematic iceberg-meltwater plumes (Fig. 9) also provide an approximation of the oceanic regions that transport less moisture to the ice sheets during different parts of the 23,000-year cycle. For minimal winter sea-ice extent during ice growth, storm tracks (Fig. 9, arrow 1) can curve to the northwest into the Labrador Sea and nourish the Laurentide Ice Sheet (Fig. 3), as well as all other circum-Atlantic ice masses.

For modest levels of melt-product flow to the ocean, the winter sea-ice extent and ineffective summer heat storage will cut off a significant part of the moisture transport from the Atlantic to the Laurentide Ice Sheet (Fig. 9, arrow 2). This configuration may, however, permit a continued or even increased meridional moisture flow northward (Fig. 9, arrow 2) through the ice-free east-central Atlantic to the Greenland, Scandinavian, and high-Arctic ice masses. Thus, a deglaciation of only moderate

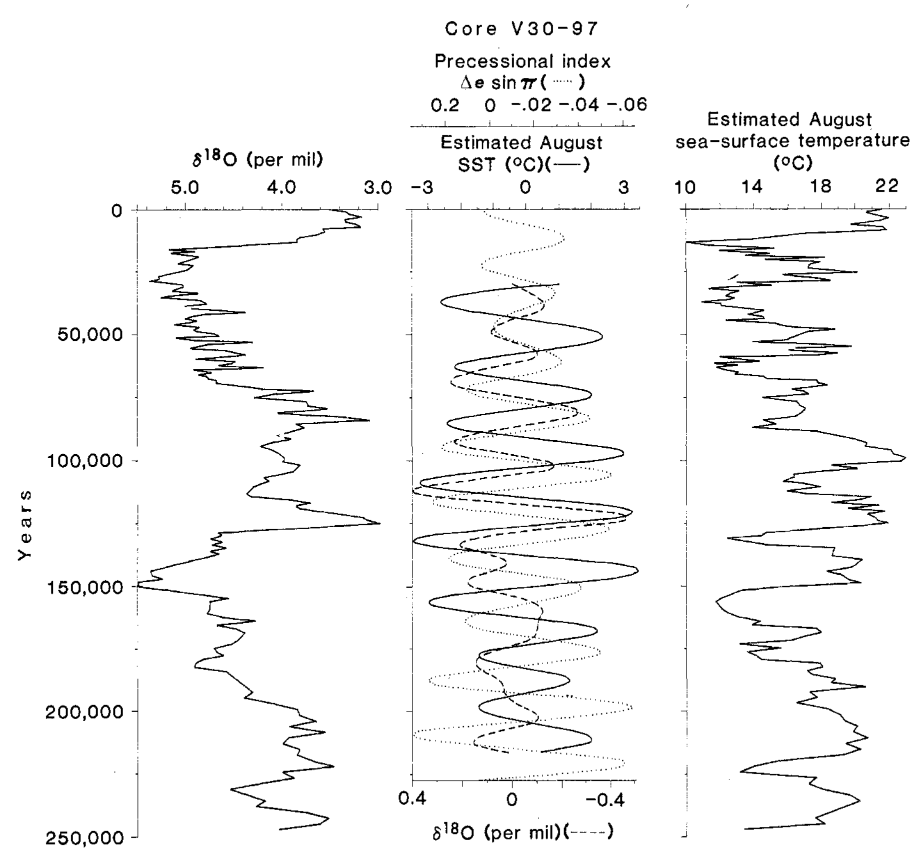

Fig. 8. (Left) Downcore $\delta^{18}O$ record in core V30-97 with time. (Right) Downcore sea-surface temperature (SST) records in V30-97. (Center) Comparison of filtered 23,000-year components of $\delta^{18}O$ and SST against the precessional index $\Delta e \sin \pi$ (*29*). Negative peaks in precession index correspond to 21 June summer insolation maxima in the Northern Hemisphere. Summer insolation leads $\delta^{18}O$ (ice volume) by 3000 ± 1500 years and SST by 9000 ± 1500 years at the 23,000-year periodicity; $\delta^{18}O$ leads SST by 6000 ± 1500 years, or one-quarter wavelength.

intensity might still permit a significant transport of moisture to high latitudes.

For the strongest intensities of melt-product flow, the winter sea-ice limit reaches to 50°N across the entire subpolar Atlantic (Figs. 4 and 9). This effectively denies significant North Atlantic moisture to all ice sheets, including even the relatively accessible Scandinavian ice mass. Winter storms (Fig. 9, arrow 3) are deflected east-southeastward toward southern Europe and the Mediterranean.

Other Oceanic Feedback Processes

In addition to the three moisture feedback loops, we have also mentioned the positive feedback from rising sea level and accelerated iceberg calving (*20*). This feedback mechanism (Fig. 5) amplifies whatever other factors are causing ice-sheet disintegration. If precessional insolation changes cause periodic (23,000-year) melting on the southern margins of mid-latitude ice sheets, the resulting changes in sea level will activate the ice-calving feedback loop at a 23,000-year cycle (*20*). This process forms a fourth major feedback mechanism for accelerating ice disintegration at the 23,000-year Milankovitch forcing period.

Finally, the feedback links in Fig. 5 are interconnected. Faster calving of icebergs not only results in ocean chilling and sea-level rise, but also enhances the meltwater layer. Decreased moisture results in a faster net rise of sea level, which accelerates the mechanical attack. Thus changes in one process are transmitted to the others.

Discussion

The data suggest a major problem with the proposal that ice-sheet melt products in part control the response of the mid-latitude North Atlantic. There is relatively weak 23,000-year power in published isotopic spectra (*6, 7*); in contrast there is a very strong 23,000-year peak in North Atlantic sea-surface temperature spectra (Fig. 7). The isotopic record implies minimal 23,000-year variations in global ice volume. We interpret the sea-surface temperature record as resulting in part from large 23,000-year variations in the volume of the dominant Laurentide Ice Sheet. How could the inferred changes in the Laurentide Ice Sheet fail to be more prominent in the global ice-volume signal?

On the basis of spectral evidence, part of the answer appears to be found in a second comparable mystery: the unexplained dominance of 100,000-year power in the isotopic spectra. Hays *et al.* (*6*) noted that orbital eccentricity modulates the amplitude of the precessional cycle at a 100,000-year period, among others, but that this power cannot appear as a linear physical response to variations in tilt or precession. They suggest that the 100,000-year power can only appear as some kind of nonlinear response of the earth's climate to orbital forcing.

Imbrie and Imbrie (*8*) explored the nonlinear response arising from a faster time constant for ice decay than for ice growth. They based this approach in part on the fact that the 100,000-year ice-volume ($\delta^{18}O$) cycles are not irregular sine waves like the eccentricity curves but tend toward a sawtooth shape (*36*). Much of the 100,000-year power is thus a

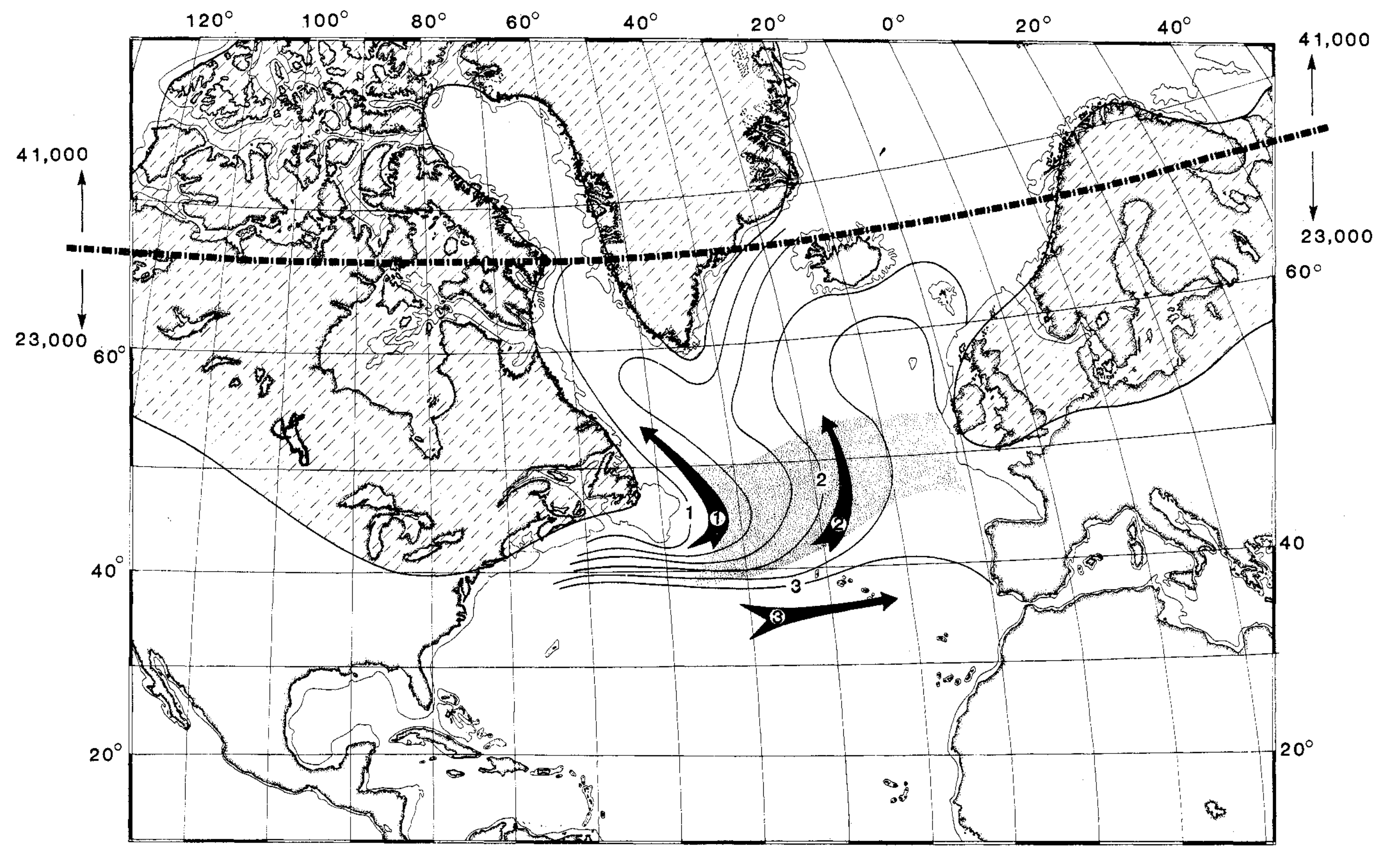

Fig. 9. Schematic representation of summer iceberg and meltwater plumes formed at different times during 23,000-year cycles of fluctuating outflow from the Laurentide Ice Sheet. Latitude 65°N separates the area dominated by the 41,000-year (tilt) frequency of summer insolation change from the region dominated by 23,000-year (precession) frequency. Stipple marks outline the region of maximum deposition of ice-rafted debris (*21*), with iceberg-meltwater plumes drawn in this general area. Largest plume extents correspond to maximum iceberg-meltwater influx. Winter sea-ice limits roughly match the meltwater plume limits. Arrows show inferred winter storm tracks and moisture transport along winter sea-ice limit during melt-product outflow at minimal (arrow 1), moderate (arrow 2), and maximum (arrow 3) levels.

consequence of the rapid (10,000-year) terminations that abruptly end longer cycles of erratic glacial buildup. Imbrie and Imbrie concluded that the nonlinearities involved in rapid ice decay can create 100,000-year power, in effect by transferring power from the 23,000-year precessional cycle. This argument suggests a solution to our problem.

The 100,000-year power in ice-volume curves is associated with the rapid glacial terminations. Because terminations begin when ice sheets are at their most southerly extent, we follow Broecker and van Donk (*36*) by inferring that the low-latitude precessional insolation maxima over the volumetrically dominant Laurentide Ice Sheet (Figs. 2 and 9) must provide the strongest initial physical impetus toward deglaciation. In contrast, the higher latitude ice sheets are smaller and the 41,000-year obliquity insolation maxima are considerably weaker (in absolute value). In effect, we suggest that the terminations can be regarded mainly as 23,000-year deglacial transitions amplified into unusually large step-function offsets.

The question of why deglaciation is more rapid than ice growth is complex, but we suggest that the oceanic feedback loops (Fig. 5) may provide part of the answer. These four feedback processes have a common attribute that is important to the timing of major deglaciation: the larger the areal extent and volume of Northern Hemisphere ice are at the time of ice disintegration, the stronger the feedback loops operate to amplify the rates of disintegration. Thus, strong precessional insolation maxima occurring during times of low-to-moderate ice volume (for example, ~ 200,000 years B.P.) (Fig. 2) do not produce deglaciations as rapid as terminations; modest precessional insolation maxima following the buildup of a large volume of ice can produce terminations such as termination III at 250,000 to 245,000 years B.P. (Fig. 2).

Thus, terminations should occur on the first precessional maximum after very large volume glacial buildup. Between terminations, the isotopic record shows a general drift toward increased ice volume, punctuated by numerous oscillations at the precessional and obliquity frequencies. This general ice-volume increase is presumably attributable to average summer insolation values that are considerably lower than those during terminations (*5*). The individual oscillations are attributable to specific summer insolation minima and maxima at the higher orbital frequencies (*8*).

We conclude that much of the 100,000-year power anomalously present in $\delta^{18}O$ records of ice volume may represent displaced 23,000-year power generated in large part by precessional forcing of the Laurentide Ice Sheet. This in part resolves the discrepancy between our proposal and the lack of 23,000-year power in isotopic (~ ice-volume) spectra.

We also note substantial 23,000-year variations in the North Atlantic sea-surface temperature response during portions of the record for which there are small or negligible variations in the isotopic record (for example, 60,000 to 30,000 years B.P.) (Fig. 8). This can be reconciled with our interpretation of the record only if (i) global (and Laurentide) ice volume actually varied at these times with more 23,000-year power than is evident in these parts of the isotopic record, or (ii) the Laurentide Ice Sheet responded to the 23,000-year forcing period at middle latitudes, while other higher latitude ice masses did not respond or responded in opposition.

The first situation could arise if the portion of the $\delta^{18}O$ record that reflects factors other than global ice volume (*2*) varies in partial opposition to the 23,000-year Laurentide ice-volume component. It could also happen if the ice formed at middle latitudes were less enriched in ^{16}O than that at high latitudes and consequently left a weaker imprint on the isotopic record (*37*). The second explanation could apply if the Laurentide ice responded with modest intensity to the smaller 23,000-year forcing period, but with the melt-product outflow concentrated only in the western North Atlantic (somewhere between plumes 1 and 2 in Fig. 9). In this outflow configuration, it would be possible to maintain a significant moisture flux northward to high latitudes in winter (Fig. 9, arrow 2), thus tending to counteract the Laurentide signal in the global ice-volume response.

In any case, geological evidence on land (*38*) shows substantial changes in the southern limit of the Laurentide Ice Sheet during the interval marked by minimal oxygen isotopic changes (60,000 to 30,000 years B.P.) (Figs. 2 and 6). These areal fluctuations imply significant oscillations in Laurentide ice volume and hence in outflow of icebergs and meltwater to the North Atlantic.

Finally, changes in oceanic delivery of heat northward from low latitudes may also contribute to the response of the high-latitude North Atlantic Ocean at the precessional frequency. This factor is implicated in the correlation between the emergence of the Panamanian Isthmus and the inception of Northern Hemisphere glaciation, both at 3.1×10^6 years B.P. (*39*). The correlation in time implies a causal link in which diversion of equatorial flow intensified the western boundary current flow in the Gulf Stream and thus delivered the warm, saline Atlantic water to middle and high latitudes that was needed in conjunction with changes in orbital insolation to fuel ice-sheet growth (*40*).

Heat transport from low latitudes is also indicated by the phasing of the 23,000-year sea-surface temperature record in core V30-97 (Figs. 6 and 7) with summer (November) insolation in the Southern Hemisphere. This suggests periodic variations in cross-equatorial heat flow in the Atlantic Ocean.

Because the ocean generally moderates atmospheric fluctuations along the coasts of continents today, many climatologists assign this role to the ocean during long-term climatic fluctuations on the scale of glacial cycles (10^4 to 10^5 years). This view ignores an oceanic role that appears to be an essential feature of ice-age cycles. On the longer time scale, the North Atlantic Ocean may amplify, rather than moderate, ice-volume changes driven by insolation. The change from a primarily negative feedback role in the short term to a positive feedback role in the long term is in part due to the action of oceanic feedback processes (Fig. 5).

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28. Almost identical sea-surface temperature spectra for this core are produced by use of the late Quaternary chronologies described by Shackleton and Opdyke (*2*), Hays *et al.* (*6*), and J. J. Morley and J. D. Hays (*Earth Planet. Sci. Lett.* in press). The chronologies are similar back to 250,000 years B.P., with the isotopic stage boundaries agreeing to within a few thousand years.
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