# 47. LEG 94 PALEOENVIRONMENTAL SYNTHESIS[1]

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**ABSTRACT**

Over 3000 m of rapidly deposited sediments with well-preserved microfossils were cored during DSDP Leg 94. High-resolution studies of these sediments have only begun. Major paleoenvironmental results to date reported in this volume include the following: (1) moderately large-scale surface-ocean changes in the east-central North Atlantic apparently began at around 3.3 Ma, in response to small fluctuations in ice-sheet volume or to other factors, such as closing of the Panamanian Isthmus or opening of the Bering Straits; (2) the 2.47 Ma date of initiation of ice rafting and large-scale Northern Hemisphere glaciation found at 56°N on Leg 81 also applies to Northeast Atlantic sites farther south to 37°N, the effective southern limit of Pleistocene ice rafting; (3) the basic rhythm of response of the North Atlantic Ocean changed from the 41,000-yr. orbital tilt period during the latest Pliocene and early Pleistocene (2.47–0.735 Ma) to the 95,000-yr. orbital eccentricity period during the later part of the Pleistocene (0.65–0 Ma), due to similar changes in the fluctuations in volume of surrounding Northern Hemisphere ice sheets.

## INTRODUCTION

The primary goal of Deep Sea Drilling Project Leg 94 was to retrieve material for paleoenvironmental studies of the late Neogene North Atlantic Ocean. For this purpose we cored six sites in a north–south transect from 53°N to 37°N in the carbonate-rich sediments along the upper and middle flanks of the Mid-Atlantic Ridge (Fig. 1). All sites have the high sedimentation rates (20–75 m/m.y.) necessary for high-resolution paleoenvironmental work.

The specific precruise paleoenvironmental objectives were divided into events occurring prior to, and subsequent to, the onset of large-scale Northern Hemisphere glaciation at about 2.47 Ma; these are listed in order of decreasing age.

For the record prior to 2.47 Ma, the main Leg 94 paleoenvironmental objectives were to determine the response of the North Atlantic to: (1) increases in volume of the Antarctic ice sheet at 14 and 6.5 Ma; (2) the closing and reopening of Atlantic/Mediterranean connections at the end of the Miocene (6–5 Ma); (3) the closing of the Panamanian Isthmus between 4.5 and 3 Ma; and (4) initial mountain valley glaciation on Iceland and in the Sierra Nevada Mountains at 3.4 to 3.1 Ma.

Three other objectives were linked to the interval of major Northern Hemisphere glaciation in the late Pliocene and Pleistocene after 2.47 Ma: (1) When did large-scale southward swings in the polar front begin and with what relation to the first large-scale glaciations? (2) Did development of strong 100,000-yr. cycles in ice volume at roughly 0.8 Ma coincide with similar changes in rhythm and amplitude of polar-front movements? (3) What was the rhythm of response of the northern subtropical gyre south of the polar front as the ice-volume rhythms changed?

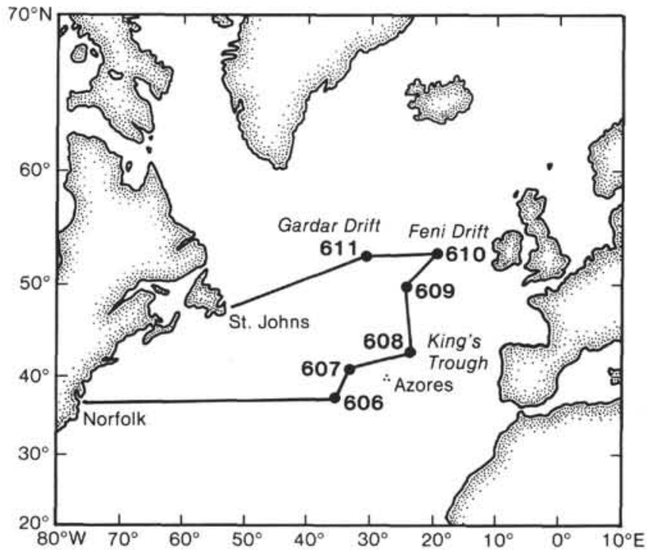


Figure 1. Location of the six Leg 94 sites.

Subsidiary objectives involving or immediately related to paleoenvironmental work included: development of high-quality paleomagnetic stratigraphy, including definition of very short events; generation of long, detailed records of oxygen isotopic and carbon isotopic signals; investigation of the detailed Neogene history of $CaCO_3$ and silica preservation and dissolution; studies of the ice-rafted and windblown terrigenous fractions; and study of the deep-water circulation history, using stable isotopes and benthic foraminiferal assemblages.

Two years have passed since the Leg 94 cruise ended; the last chapters to be included the Leg 94 *Initial Reports* have now been accepted for publication. Because state-of-the-art paleoenvironmental work requires sampling at a small scale (about 5 kyr., or kiloyears) to avoid signal aliasing at the primary orbital frequencies (20–100 kyr.), the chapters included in this volume are only a prelude to those that will ultimately be published. Nev-

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ertheless, we can already report substantial progress toward many of our objectives.

## OUTLINE OF MAJOR RESULTS

The principal paleoenvironmental findings from Leg 94 sediments of post-Oligocene age are summarized in Figure 2. Also shown in Figure 2 are the major global climatic and tectonic changes relevant to Leg 94 results.

### Early/Middle Miocene

Lower-middle Miocene sediments were only recovered at Sites 608 and 610, and few Leg 94 studies of these intervals have been completed. Baldauf (this volume) recognized a decrease in diatom preservation in the lower-middle Miocene at Site 610 and an apparent short hiatus spanning the lower/middle Miocene boundary at Site 406. Although the apparent hiatus most likely reflects an increase in bottom-water circulation, the decrease in diatom preservation could reflect either a decrease in surface-water productivity or increased dissolution of biogenic silica. Thomas (this volume) noted episodes of minor turnover in benthic foraminiferal faunal composition at 19.2 to 17 Ma and 15.5 to 13.5 Ma.

At Site 608, Miller et al. (this volume) found carbon isotopic evidence that deep waters in the eastern North Atlantic were isolated from the well-ventilated western basin from prior to 24 Ma until about 15 Ma. The decreasing isolation of the eastern basin at about 15 Ma was ascribed to increased advection of deep waters from the north across the subsiding Iceland-Faeroe portion of the Greenland-Scotland Ridge.

### Late Miocene

Upper Miocene and younger sediments from Leg 94 have been more intensively studied. Results suggest that major changes in the deep circulation occurred between 7.0 and 5.5 Ma.

Hooper and Weaver (this volume) interpreted particularly intense $CaCO_3$ dissolution at Site 609 in the upper Miocene (corresponding in time to after 7.0 Ma) as indicating a strong flow of corrosive bottom water from Antarctica. They attributed the progressive decrease in dissolution during the later Miocene to an increase in the relative flow of deep water from northern sources. Murray (this volume) interpreted benthic foraminiferal assemblage changes at Sites 609 to 611 as indicating strong flow from Antarctic sources just prior to 6.0 Ma. Thomas (this volume) also found a significant turnover in the composition of benthic foraminifers from 7.0 to 5.5 Ma at a shallow site (610; ~2.5 km), but not at a deeper site (608; ~3.5 km). Keigwin et al. (this volume) noted that the $\delta^{13}C$ values registered by *Planulina wuellerstorfi* at Site 552 in the latest Miocene suggest that North Atlantic Deep Water continued forming during the Messinian.

As for surface waters, Hooper and Weaver (this volume) interpreted the predominance of left-coiling *Neogloboquadrina pachyderma* (sinistral coiling; the unencrusted form) from 6.5 to 5.6 Ma at Site 609 as evidence of a minimum temperature of cool surface waters at 6.2 Ma. Large oscillations between the sinistrally and dextrally coiling *N. pachyderma* populations from 5.6 to 4.8 Ma were similarly interpreted as major oscillations in sea-surface temperature, with an apparent periodicity around 125 kyr. at the 20-kyr. sampling interval. This use of coiling directions of *N. pachyderma* in the upper Miocene for paleoclimatic interpretations follows earlier work by Poore and Berggren (1975), but Hooper and Weaver also pointed out that Kennett and Vella (1975) have argued that this parameter was not an indicator of surface-ocean conditions until the early Pliocene. As a further caution, Keigwin (1978) has reported "enigmatic" abundances of left-coiling *N. pachyderma* in Caribbean and eastern equatorial Pacific sediments during the latest Miocene and early Pliocene.

Keigwin et al. (this volume) found quasiperiodic fluctuations in $\delta^{13}C$ records from benthic and planktonic foraminifers at Site 552. They speculated that these oscillations may have been in response to changes in continental biomass at or near orbital periodicities.

### Early Pliocene (5.3-3.5 Ma)

There is evidence that significant bottom-water changes occurred during the early Pliocene. At Site 609, Hooper and Weaver (this volume) noted a transition at roughly 5.2 Ma from weakly corrosive bottom waters interpreted as originating in the North Atlantic to more strongly corrosive waters presumably from an Antarctic source, and then back to less corrosive (northern-source?) waters at roughly 4.5 Ma. This interpretation agrees only in part with the benthic foraminiferal assemblage data of Murray (this volume), who inferred a maximum in Antarctic-source bottom waters from 4.6 to 3.2 Ma. Thomas (this volume) did not find changes in benthic foraminiferal assemblages at Sites 608 and 610 during this interval (see also Kidd and Hill, this volume).

Keigwin et al. (this volume) noted two brief $\delta^{18}O$ maxima at the beginning of the early Pliocene at Site 552, with both the covariance between the planktonic and benthic $\delta^{18}O$ records and the large amplitudes of the signals (0.5‰) indicating brief increases in continental ice volume at 5.5 and at 5.0 Ma. These pulses fall within the interval of large oscillations in circulation patterns of North Atlantic surface-water, inferred by Hooper and Weaver (this volume) from faunal data at Site 609. From planktonic foraminiferal assemblages at Site 609, Hooper and Weaver (this volume) inferred that a uniformly warm interval occurred in the early Pliocene after 4.8 Ma.

### Late Pliocene (3.5-1.66 Ma)

A number of indicators point to major changes in deep-water circulation in the first half of the late Pliocene. Ruddiman et al. (this volume) found increasingly intense depth-dependent $CaCO_3$ dissolution from 3.15 to 2.5 Ma. Because increased dissolution in the Atlantic is usually ascribed to increased flow of deep waters from Antarctic sources, this evidence suggests stronger Antarctic-source flow after 3.15 Ma. In contrast, Murray (this volume) interpreted benthic foraminiferal assemblages at Sites 609 and 611 to indicate that deep waters resembling those in the eastern North Atlantic today ("NEADW") first developed at those Sites at about 3.2

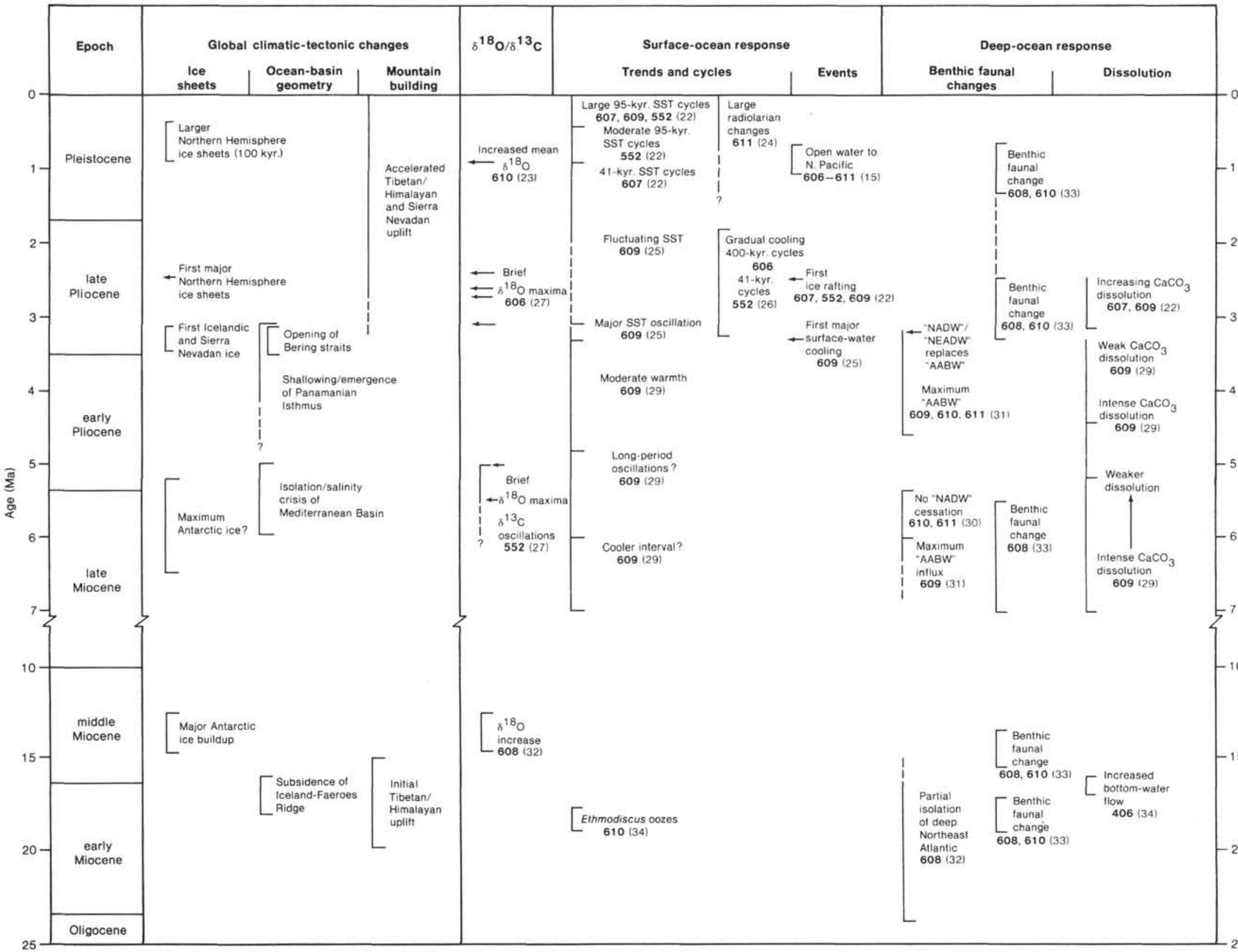


Figure 2. Synthesis of Leg 94 paleoenvironmental results and comparisons to timing of global climatic and tectonic events. Sites at which observations were made are listed; relevant chapter numbers are indicated in parentheses.

Ma. Thomas (this volume) noted a major change in benthic faunal composition at Sites 608 and 610 between 3.15 and 2.4 Ma, but did not interpret the results.

Oxygen isotopic evidence indicates some combination of increased ice volume and deep-water cooling between 3.1 and 2.4 Ma. Keigwin (this volume) found short $\delta^{18}O$ maxima at 3.1, 2.7, 2.6, and 2.4 Ma at Site 606, superimposed on a background trend of increasing isotopic values. He suggested that minor glaciation occurred during the earliest pulse (3.1 Ma), but that this was mostly a deep-sea cooling event. Keigwin dated the permanent $\delta^{18}O$ enrichment at 3.1 Ma, comparable to earlier estimates (Prell, 1984).

At every site on Leg 94, unequivocal evidence was found that substantial Northern Hemisphere glaciation began during the interval from 2.55 to 2.4 Ma, apparently coincident with the Gauss/Matuyama boundary at 2.47 Ma. Ruddiman et al. (this volume) noted that the effects of this important late Pliocene event, reported by Berggren (1972) and Poore and Berggren (1975) and later accurately dated by Backman (1979) and Shackleton et al. (1984), thus influenced the entire sector of the eastern and central North Atlantic reached by large-scale ice rafting in the last half of the Pleistocene (as far south as 37°N). Latouche et al. (this volume) found that detritus of glacial origin became the dominant component of the clay-mineral fraction at all Leg 94 sites after 2.4 Ma.

Aside from the initiation of large-scale ice-rafting, surface-water changes in the late Pliocene are somewhat difficult to interpret. At Site 606 (37°N), Backman and Pestiaux (this volume) found a gradual decline in discoaster abundance from 3.25 Ma to the time of total discoaster extinction at 1.85 Ma. They attributed this decline to a prolonged cooling of northern subtropical waters in the North Atlantic. Superimposed on this decline were 400-kyr. fluctuations in discoaster abundance. An earlier study had shown that the same decline at Hole 552A at 56°N contained significant power at a period of 41,000 yr. (Backman et al., in press).

Raymo et al. (this volume) interpreted planktonic foraminiferal trends as indicating a major cooling from 3.4 to 3.2 Ma at Site 609 (see also Ehrmann and Keigwin, this volume). This was followed by an interval of intermediate and oscillatory conditions between 3.2 and 2.47 Ma, with periodic surface-water oscillations to conditions cooler than today but not as cold as during the glaciations of the late Pleistocene. They also found substantial relative abundance changes in the planktonic foraminiferal fauna near the beginning of major Northern Hemisphere glaciation at 2.47 Ma, with most changes suggesting a cooling. Climatic interpretations of planktonic foraminifers during this interval are complicated by the evolutionary overturn of prominent species. The rest of the late Pliocene foraminiferal fauna at higher latitudes remains difficult to interpret because of no-analog faunas (heightened abundances of some species relative to their percentages today and/or the virtual absence of other species that are prominent today).

### Pleistocene

A large turnover of benthic fauna that took place during the Pleistocene at Sites 608 and 610 between 1.3 and 0.7 Ma was noted by Thomas (this volume). Otherwise, most Leg 94 studies of the Pleistocene were focused on the surface-water indicators.

There are some disparities among different paleoceanographic indicators about the basic sense of Pleistocene surface-water changes. Morley (this volume) found that the radiolarian *Cycladophora davisiana*, thought to indicate extensive developement of sea ice and salinity-stratified surface waters in the polar seas, held to the same basic range of abundance fluctuations through the entire Pleistocene at Site 611. This suggests little change in the amplitude of glacial–interglacial surface-water fluctuations between the Brunhes and the Matuyama chrons, although Morley noted that radiolarians were not present through large portions of the early Pleistocene.

On the other hand, Ruddiman et al. (this volume) found a significant increase in the amplitude of fluctuations in planktonic foraminiferal faunas and in estimated sea-surface temperature (SST) relative to the earlier Pleistocene record at Site 607 (41°N) after 1.0 to 0.8 Ma. Ruddiman et al. (in press) also found a major increase in fluctuations in foraminiferal abundances and SST between 1.0 and 0.4 Ma at Site 552 (56°N).

Other evidence strongly supports a major surface-water change between 1.0 and 0.4 Ma. The increase in amplitude of percent $CaCO_3$ fluctuations noted by Ruddiman et al. (this volume) at Sites 607 and 609 during the last million years indicates that North Atlantic ice rafting increased and expanded southward. There may also have been greater suppression of productivity and stronger glacial dissolution during the glacial maxima. These changes agree with oxygen isotopic analyses made at very broad time intervals at Site 610 (Jansen and Sejrup, this volume); these data repeat trends established by previous more detailed $\delta^{18}O$ studies showing larger ice-volume maxima since 0.9 Ma (Shackleton and Opdyke, 1973; Prell, 1982). On balance, the evidence suggests higher-amplitude Northern-Hemisphere climatic fluctuations during the latest Matuyama and Brunhes than in the earlier Pleistocene.

Baldauf (this volume) observed diatoms of the *Denticulopsis seminae* group only in middle Pleistocene sediments at Sites 607, 609, 610, and 611. This diatom species is today characteristic of the North Pacific and the Bering Sea. Its presence suggests communication between the Bering Sea and North Atlantic through the Arctic Ocean during the middle Pleistocene (1.07–0.73 Ma), possibly accompanied by lower salinity, lower temperature and changes in nutrients in the North Atlantic.

The rhythms of the North Atlantic sea-surface temperature responses also changed during this mid-Pleistocene interval. For the last 0.45 m.y. of the Pleistocene, the basic tempo of fluctuations in sea-surface temperatures in the polar gyre north of 45°N has been 95,000 yr.; prior to this time, the 95,000-yr. SST rhythm was weaker by as much as 75% (Ruddiman et al., in press). Oxygen isotopic records show a similar evolution of 100,000-yr. power during this time (Pisias and Moore, 1981; Imbrie, 1985). The North Atlantic SST signal at 95,000 yr. is directly in phase with the $\delta^{18}O$ signal, implying ice-sheet control of the surface-ocean response (Ruddiman et al., in press).

During the last 250,000 yr., the region of Site 607 (41°N) responded with very strong SST power near 100,000 yr. and also at 23,000 yr. (Ruddiman and McIntyre, 1984). Initial spectral analysis of full Pleistocene records of estimated SST and percent $CaCO_3$ at Site 607 has just been completed using a composite record similar to that shown in Ruddiman et al. (this volume) but incorporating accurately controlled between-hole splices to span gaps and disturbances at core breaks (Ruddiman et al., in press). This analysis showed two major results: (1) 100,000-yr. power was negligible in both records prior to 1.0 Ma, amounting to no more than 5 or 10% of its absolute late Pleistocene value; (2) 41,000-yr. power was as dominant from 2.0 to 1.0 Ma as 100,000-yr. power has been during the Brunhes Chron. The "middle" Pleistocene (ca. 0.9–0.6 Ma) thus marks a fundamental shift in response of the North Atlantic from one orbital periodicity to another.

The use of transfer functions on planktonic foraminiferal data from Leg 94 sites at and north of 50°N is limited to the Pleistocene. Prior to 1.7 Ma, *N. pachyderma* (s.) was very rare at these sites, despite the presence of abundant ice-rafted debris and $CaCO_3$ minima suggestive of cool surface waters; this anomalous condition developed again between 1.25 and 1.1 Ma (Ruddiman et al., this volume; Raymo et al., this volume). At sites north of 50°N, these intervals were marked by abundances of *Globorotalia inflata* and/or *N. pachyderma* (d.) at "no-analog" levels, that is, at percentages greater than those observed in any North Atlantic surface sediments. Under these conditions, transfer functions cannot be used.

At the more southern latitude of Site 607 (41°N), where *N. pachyderma* (s.) is a less important member of the Pleistocene fauna even during glaciations, the transfer functions basically remain useful back through at least the 1.9 m.y. record analyzed to date. Slightly higher estimated SST values occur at the intervals where *N. pachyderma* (s.) is absent, and this apparent bias appears to impart a low-frequency wave to the SST trend (Ruddiman et al., this volume). This suggests a marginal impact from the no-analog condition that is severe farther north.

## DISCUSSION AND CONCLUSIONS: LINK OF LEG 94 RESULTS TO GLOBAL CHANGES

We now reexamine the questions posed in the introduction to this chapter. We summarize briefly the progress to date in answering these questions and note likely directions for ongoing and future research in this area.

***What was the response of the North Atlantic to increases in Antarctic glaciation around 14 and 6.5 Ma?***

For both these intervals, there are North Atlantic responses suggestive of possible links to changes in the Antarctic environment. Shackleton and Kennett (1975) interpreted the earlier increase in oxygen isotope values at about 14 Ma as indicating the formation of large ice sheets on Antarctica, although Savin et al. (1975) emphasized cooling of deep waters in contributing to the middle Miocene isotopic shift. Miller et al. (this volume) dated this isotopic change at 14.6 to 12.6 Ma by correlation to the paleomagnetic record (Clement and Robinson, this volume). Time-correlative events identified in the northeast Atlantic include: increased bottom-water velocity and/or corrosiveness in the late Miocene (Hooper and Weaver, this volume); and an end to the partial isolation of eastern North Atlantic deep water in the middle Miocene (Miller et al., this volume). A change in benthic foraminiferal composition preceded this circulation change by about 2 m.y. (Thomas, this volume; Murray, this volume); another faunal change appears to be about coeval (Fig. 2).

All of these changes could be connected to changes in formation of Antarctic-source deep waters associated with the major increase in Antarctic glaciation. At this point, however, proposed connections between Antarctic ice volume, deep-water temperatures, and North Atlantic surface-water responses for the middle Miocene must be highly speculative. The atmospheric circulations of these two polar regions are thermally and dynamically isolated from each other (Manabe and Broccoli, 1984), and the only plausible connections from the circum-Antarctic to the Northern Hemisphere are via the deep-water circulation (with speculative linkages) or via atmospheric $CO_2$ levels. In addition, the exact timing of all these middle Miocene phenomena remains just uncertain enough to make it difficult to use their relative timing for evaluating critical cause-and-effect relationships.

An early stable isotope study suggested that there was a major increase in Antarctic ice volume in the latest Miocene (Shackleton and Kennett, 1975). This was thought to have been associated with expansion of Antarctic polar waters and acceleration of the Antarctic Circumpolar Current (Cieselski et al., 1982) and with pronounced regional cooling trends around the Pacific (Ingle, 1967; Kennett and Vella, 1975). Higher-resolution studies, however, reveal only brief $\delta^{18}O$ events, and not a permanent baseline shift in upper Miocene isotopic values (Keigwin, 1979; Keigwin et al., this volume). Detailed study of the 6.5 to 5.0 Ma interval at Site 552 shows that the important $\delta^{18}O$ events are dated at 5.5 and 5.0 Ma (Keigwin et al., this volume).

***What was the response of the North Atlantic to closing and reopening of connections with the Mediterranean at the close of the Miocene during the Messinian?***

Several substantial changes shown in Figure 2 have been detected in Leg 94 records of Messinian age: brief $\delta^{18}O$ maxima indicative of deep-water cooling and/or increased global ice volume (Keigwin et al., this volume); quasiperiodic changes in $\delta^{13}C$, which may be indicative of changes in continental biomass at orbital frequencies (Keigwin et al., this volume); large oscillations in planktonic foraminiferal coiling direction at periods near 125,000 yr. (Hooper and Weaver, this volume); changes in benthic foraminiferal assemblages (Murray, this volume; Thomas, this volume); and indications of weaker $CaCO_3$ dissolution attributed to a decreased volume of Antarctic-source deep water after about 6 Ma (Hooper and Weaver, this volume) (see also Masson and Kidd; and Kidd and Hill, this volume).

Hooper and Weaver (this volume) suggested that the Messinian isolation of the Mediterranean may be connected to late Miocene fluctuations in *N. pachyderma* coiling directions between 6 and 5 Ma. As noted above, the environmental significance of this parameter in the Miocene is still unclear. This is an important topic for more detailed work in the future, as are the cyclical $\delta^{13}C$ changes (Keigwin et al., this volume).

Several theoretical connections between the Mediterranean isolation and other aspects of global climate during the Messinian have previously been proposed in the literature. Two of these link Antarctic glaciation and Mediterranean isolation, but in inverse senses. Increases in the volume of Antarctic ice could lead to isolation of the Mediterranean via the link of sea level (Adams et al., 1977; Loutit and Keigwin, 1982). On the other hand, isolation and dessication of the Mediterranean could cause Antarctic glaciation by another cause-and-effect sequence: reducing circum-Antarctic salinity, enhancing Antarctic upper-water stability, promoting Antarctic sea-ice formation, increasing mean circum-Antarctic albedo, and decreasing circum-Antarctic temperatures (Ryan et al., 1974). Leg 94 evidence to date (Keigwin et al., this volume) does not bear directly on these ideas, other than to support the existence of short positive $\delta^{18}O$ pulses in the late Miocene that could represent as much as 60-m temporary decreases in global sea level, if all the $\delta^{18}O$ change represents ice accumulated at 0.1‰ per 10 m of sea level. On the other hand, none of these pulses requires ice-volume changes, because temperature fluctuations of bottom waters could absorb all the variation without freezing the deep ocean.

Two other theories link Mediterranean dessication to deep-water processes. Blanc and Duplessy (1982) suggested (based on low-resolution $\delta^{13}C$ signals) that deep-water formation might cease in the North Atlantic during Mediterranean dessication because the lowered North Atlantic salinities would stabilize the water column and suppress overturn. However, additional carbon isotopic data indicate that formation of deep water continued uninterrupted during the Messinian without the Atlantic having an open connection with the Mediterranean (Keigwin et al., this volume). Ryan et al. (1974) proposed that extraction of excess amounts of sulphates and carbonates by the Messinian evaporite deposition in the Mediterranean would lead to severe dissolution elsewhere in the world ocean. At least on the local scale of Site 609 in the North Atlantic, however, Hooper and Weaver (this volume) find evidence suggesting the opposite trend and thus arguing against both theories.

***What was the response of the North Atlantic to closing of the Panamanian Isthmus at 4.5 to 3 Ma? What was the response of the North Atlantic to climatic coolings evidenced by early glaciation on Iceland and in the Sierra Nevada Mountains at 3.4 to 3.2 Ma? When did major southward swings of the polar front begin, and were they synchronous with major ice-sheet fluctuations?***

We combine these precruise objectives because they involve approximately the same interval of time and because they may involve causally related phenomena. We rephrase these questions as follows: Were there large-scale oscillations of North Atlantic surface waters before major ice-sheet fluctuations began at 2.47 Ma? If so, what caused them?

Late Pleistocene fluctuations in position of the polar front were synchronous with, and apparently controlled by, the size of Northern Hemisphere ice sheets (Ruddiman and McIntyre, 1984). The specific mechanism of ice-air-ocean interaction appears to be cold, strong westerly winds that blew from the northern flank of the North American (Laurentide) ice sheet out into the North Atlantic Ocean, extracting the little heat available from those high-latitude waters (Manabe and Broccoli, 1984).

Because of this well-defined link, it might be anticipated that the sudden initiation of relatively large-scale ice rafting at 2.47 Ma, which marks an important increase in the development of large-scale Northern Hemisphere ice sheets, would cause the first substantial cooling of North Atlantic surface waters and the first major polar-water advance. Earlier surface-water fluctuations, however, clearly occurred and require an explanation.

Poore (1981) found long-term (~1-m.y.) fluctuations in foraminiferal faunal composition in upper Miocene sediments (10–5 Ma). Leg 94 data show that large-scale oscillations in at least one planktonic foraminiferal index species (*N. pachyderma* coiling direction) occurred from 5.6 to 4.8 Ma, and may be interpreted as reflecting changes in the sea-surface response (Hooper and Weaver, this volume). The fluctuations prior to the late Pliocene have generally been sampled in less detail, and their interpretations tend to be more equivocal because of the overprint of evolutionary complications (and dissolution at Site 609). More work is needed both to verify that these earlier (pre-2.47 Ma) changes are indicative of surface-ocean responses and to test for periodicities.

Clear evidence of quasiperiodic surface-water fluctuations first occurred at 3.3 Ma. Earlier studies of planktonic foraminifers in rotary drilled cores from Leg 12 suggested a major cooling of the Labrador Sea in the late Pliocene at about 3 Ma (Berggren, 1972), but the lack of paleomagnetic stratigraphy and the drilling disturbances did not permit an accurate age estimate for this change, nor could surface-ocean oscillations be detected from these data.

Planktonic foraminiferal changes based on Leg 94 (and previous) hydraulic piston coring (HPC) pinpoint a major cool-to-warm surface-water oscillation in the late Pliocene from 3.4 to 3.2 Ma, followed by periodic (or quasiperiodic) oscillations in sea-surface response at moderate amplitudes into the Pleistocene (Raymo et al., this volume; Loubere and Jakiel, 1984; P. Loubere, personal communication, 1985). The *Discoaster* spp. abundance changes beginning at about 3.25 Ma are interpreted as indicating an oscillatory cooling trend lasting until final extinction of all discoasters in the late Pliocene at 1.85 Ma (Backman and Pestiaux, this volume).

Thus the firmest Leg 94 evidence of the beginning of quasiperiodic North Atlantic surface-water fluctuations is at about 3.3 Ma. This is also the approximate age (within radiometric and stratigraphic dating uncertainties) at which other climatic indicators place a fundamental change in Northern Hemisphere climate toward a colder state: initial mountain glaciation in the Sierra

Nevada (Curry, 1966) and in Iceland (McDougall and Wensink, 1966); and the beginning of late Pliocene cooling of the Mediterranean (Thunnell, 1979).

A change toward heavier $\delta^{18}O$ values is also evident in many oxygen isotopic records at about this time. It has been variously interpreted as indicating permanent or temporary (1) increases in ice volume in the Northern and/or Southern Hemispheres, or (2) decreases in temperature of bottom waters and high-latitude surface waters (Keigwin, this volume; Prell, 1984). This would seem to suggest a correlation between the surface-ocean cooling at 3.3 Ma and the $\delta^{18}O$ enrichment indicative of larger ice sheets or cooler bottom waters. At present, uncertainties in absolute ages and in correlations between records preclude making a firm link between the North Atlantic response and the $\delta^{18}O$ record around 3.2 Ma.

Other Leg 94 evidence (Fig. 2) also points to important changes in bottom waters during this late Pliocene interval: a major change in composition of the benthic foraminiferal fauna (Murray, this volume; Thomas, this volume); and evidence of increasing severity of depth-dependent dissolution (Ruddiman et al., this volume). We have already noted, however, that these studies disagree as to whether flow increased from Northern or Southern Hemisphere sources. Future studies of $\delta^{13}C$ in the eastern North Atlantic might help to choose among these possibilities.

In any case, it is clear that major changes in surface and deep circulation of the North Atlantic and in climate on the surrounding continents began within a relatively short interval in the late Pliocene around 3.4 to 3.1 Ma. These climatic changes must have been due either to: (1) changes in volume of small ice sheets (on remote northern land masses) that did not send icebergs into the North Atlantic; or (2) factors other than Northern Hemisphere ice sheets.

The lack of ice-rafted debris in the east-central North Atlantic prior to 2.47 Ma does not rule out earlier Northern Hemisphere ice. By analogy with today, ice could exist on Greenland and locally in the Canadian Arctic without ice rafting reaching the area cored by Leg 94 or other DSDP cruises. Recent preliminary results on Legs 104 and 105 of the Ocean Drilling Project (ODP) suggest initiation of some ice rafting at least 0.5 m.y. prior to the change at 2.47 Ma. Assuming that this was not caused solely by sea ice picking up debris from continental margins, ice sheets during this earlier interval must have been very small.

It is also possible that other factors influenced the North Atlantic surface-ocean response after 3.4 Ma. Major tectonic changes (Fig. 2) occurred at roughly the same time: final emergence of the Panamanian Isthmus (Saito, 1976; Keigwin, 1978), and opening of the Bering Strait (Einarsson et al., 1967). Individual chapters in this volume have made no specific attempt to draw a causal connection between the response of the North Atlantic and these changes in solid boundary conditions of the earth's crust, despite the suggestive time equivalence. These possible linkages call for more work in the future.

***Did development of strong 100,000-yr. cycles in ice volume roughly 800,000 yr. ago coincide with similar changes in rhythm of polar front movements? What was the rhythm of oceanic response in the northern subtropical gyre south of the polar front as the ice volume rhythms changed?***

Again, we combine several interrelated questions and answers. Our major finding to date from Leg 94 and other North Atlantic HPC cores is summarized in Figure 3: the dominant rhythm of climatic variation in the North Atlantic systematically shifted from the 41,000-yr. period of orbital obliquity in the first half of the Pleistocene (prior to 1.0 Ma) to the 95,000-yr. period of orbital eccentricity in the last half of the Pleistocene. Prior to 1.0 Ma, the 95,000-yr. period was virtually absent from the surface-ocean response, whereas the 41,000-yr. period was at least three times stronger in absolute terms than in the Brunhes (Ruddiman et al., in press). During the late Pleistocene, the amplitude and relative dominance of the 95,000-yr. signal became twice as large in the last 0.4 m.y. as it had been during the previous 0.4 m.y.

In large part, the Northern Hemisphere ice-sheet response apparently evolved in a similar manner. Imbrie (1985) found that the "100K" (100,000 yr.) $\delta^{18}O$ signal also was only half as strong between 0.8 and 0.4 Ma as it has been in the last 0.4 m.y. A continuous $\delta^{18}O$ record sampled at adequate resolution prior to 1.0 Ma does not exist, but the record from Hole 552A appears to be dominated by a period near 41,000 yr. in disjunct segments at

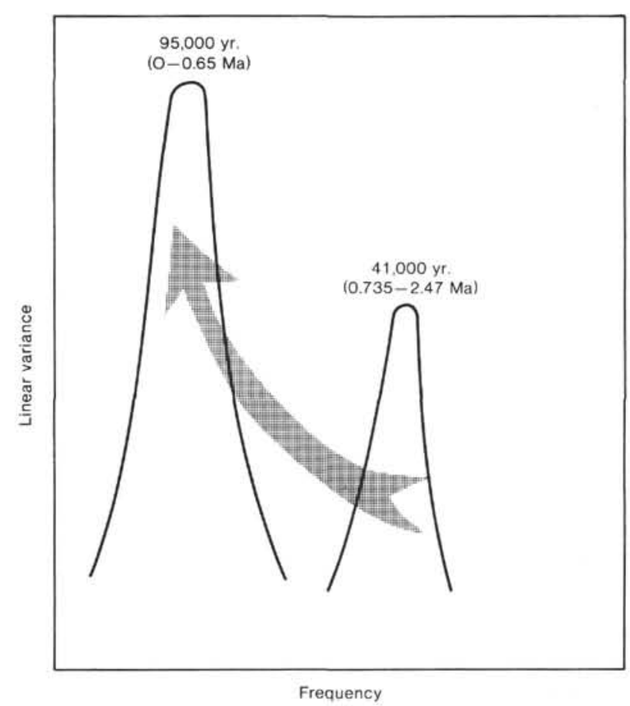


Figure 3. Pleistocene evolution of Northern Hemisphere ice sheets is indicated by showing the dominant climatic response (ice-air-ocean) in the North Atlantic region during this time. The dominant response has shifted toward longer periodicities and higher amplitudes (from Ruddiman et al., in press).

2.0 to 1.4 and 1.2 to 1.0 Ma (Shackleton et al., 1984). These correlations between the SST and $\delta^{18}O$ records appear to confirm the temporal and causal linkage between Northern Hemisphere (North American) ice sheets and the high-latitude North Atlantic SST response noted above.

Even though this evidence of rhythmic Pleistocene responses appears to support the concept that Northern Hemisphere ice sheets are driven by orbitally controlled insolation and in turn provide the dominant immediate local forcing of high-latitude North Atlantic SST, the possible presence of orbital rhythms in North Atlantic responses prior to the 2.47-Ma initiation of major glaciation remains unexplained. P. Loubere (personal communication, 1985) found an apparent periodicity near 42,000 yr. in the late Pliocene (3.0–2.5 Ma) responses of several species of planktonic foraminifers at eastern North Atlantic Site 548 (45°N). On the other hand, discoaster abundances at Site 606 (37°N) show 400,000-yr. power in the late Pliocene (Backman and Pestiaux, this volume).

As noted earlier, these earlier responses require forcing either by: (1) small Northern Hemisphere ice sheets at very high latitudes, or (2) some other part of the climate system. The fact that these SST responses occur at orbital periods indicates that unknown factors play the role of intermediaries that transfer variations in incoming insolation to the surface ocean. This is a particularly interesting problem for future research.

The changes in dominant rhythmic response of the ice sheets and the surface North Atlantic Ocean after 2.47 Ma (Fig. 3) remain unexplained. The shift from short to long periods basically indicates a change in the sensitivity of the terrestrial climate system. From among the large array of orbital forcing periods, only a small band width is selected for strong amplification at high latitudes of the Northern Hemisphere at any one time. This amplification is attributed to factors internal to the climate system (Ruddiman et al., in press).

Changes in these internal amplification processes are probably caused by tectonic changes in the solid boundary conditions at the Earth's surface that also led to the initiation of major glaciation at 2.47 Ma. Another factor that could be important in the progressive (or abrupt) deterioration of Northern Hemisphere climate is the extent of Arctic sea-ice cover. Although there is still considerable disagreement over the history of Arctic sea ice (Herman and Hopkins, 1980; Clark, 1982), major changes in Arctic sedimentation estimated to have occurred at or near 2.5 and 0.7 Ma coincide with the initiation of glaciation and with the major change in dominant North Atlantic response. This suggests a link, but it is unclear whether the Arctic response is cause or effect. Results from ODP Legs 104 and 105 may bear on this problem.

Finally, Leg 94 results have at this point not clarified the full climatic role of the surface North Atlantic Ocean. It clearly reacts to (and reflects) climatic changes on adjoining land masses. But two important questions raised by Ruddiman and McIntyre (1984) remain unanswered:

(1) What role does the North Atlantic play as a climatic feedback factor?

(2) Has the North Atlantic played an active and independent role in the evolution of Northern Hemisphere climate, including the fluctuations of the ice sheets?

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