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THE INITIATION OF NORTHERN HEMISPHERE GLACIATION

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INTRODUCTION

For well over a century, scientists have speculated that variations in the atmospheric concentration of radiatively important trace gases, such as carbon dioxide, could control the Earth's climate (e.g. Arrhenius 1896). Recent observations of significant glacial-interglacial variation in atmospheric CO_2 in ice cores put this idea on a firm observational as well as theoretical footing (e.g. Barnola et al 1987). In this century, as atmospheric CO_2 levels creep steadily upwards in response to the burning of fossil fuels and forests, the specter of manmade global climate change looms before us (IPCC 1990)—a possibility that has heightened interest in the natural variability of climate in the past. In particular, the $\sim$3°C warming predicted for the next century has stimulated interest in past warm climates, such as the Middle Pliocene around 3.0 million years (Ma) ago, an interval that has often been invoked as a "greenhouse" world (Budyko et al 1985).

In this paper, the climate transition from the warm mid-Pliocene (around 3.2 Ma) to the onset of northern hemisphere ice ages around 2.4 Ma is examined. Evidence for the initiation of significant northern hemisphere glaciation is examined as well as how this event affected climate around the globe. While the cause of individual glacial-interglacial oscillations is tied to Milankovitch variations in the Earth's orbit around the sun (e.g. Imbrie et al 1992), these insolation changes cannot account for the long-term cooling trend which culminated in northern hemisphere glaciation. In the final section of this paper, mechanisms of long-term climate change

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are examined with special emphasis on those that have been proposed to explain the late Neogene cooling of the northern hemisphere.

Throughout this paper, data and figures are presented and discussed in accordance with the magnetic time scale of Berggren et al (1985). Although a revised magnetic time scale (Shackleton et al 1990, Cande & Kent 1992) is now available to the geologic community, almost all of the studies discussed in this review were published using the Berggren et al time scale, hence its adoption here. For the interval between 2 and 3 Ma examined below, one can approximate the more recent magnetic time scale by adding 6% to all ages. Isotopic stage designations, some of which are indicated on Figures 1 and 3, are obviously not affected by the choice of time scales. They refer to specific features of the oxygen isotope record, features whose age depends on the time scale being used. In this paper, the isotopic stage designations of Raymo et al (1989) are used.

EVIDENCE FOR NORTHERN HEMISPHERE GLACIATION

The general outline of Pliocene-Pleistocene climate history has been known since the 1970s. This knowledge derives primarily from the measurement of oxygen isotope ratios ($^{18}O/^{16}O$) in the shells of calcitic foraminifera. When ice sheets grow on land, ^{16}O, the light isotope of oxygen, is preferentially extracted from the oceans and concentrated in continental ice sheets. This causes the ocean isotopic ratio $\delta^{18}O$:

$$\delta^{18}O = \frac{(^{18}O/^{16}O)_{sample} - (^{18}O/^{16}O)_{standard}}{(^{18}O/^{16}O)_{standard}} \times 1000$$

to get correspondingly heavier, or more positive. In addition, a temperature-dependent fractionation between water and calcite also causes the $\delta^{18}O$ value of calcite to increase as ocean temperatures cool. Thus, glacial climates are associated with more positive $\delta^{18}O$ values while more negative values are associated with warmer interglacial climates (Figure 1). For a detailed review of this methodology see Mix (1987).

Shackleton & Opdyke (1977) applied this oxygen isotope technique to the question of when the Ice Ages started. Using an extremely low sedimentation rate piston core (<1 cm/Kyr) from the equatorial Pacific Ocean, these authors were able to generate a $\delta^{18}O$ record that extended back to the Gauss magnetic chron (2.47–3.40 Ma). Although the V28-179 record was of poor quality and resolution due to the low sedimentation rate, intense bioturbation, and low carbonate content of the core, many of Shackleton & Opdyke's conclusions are still valid today. They proposed

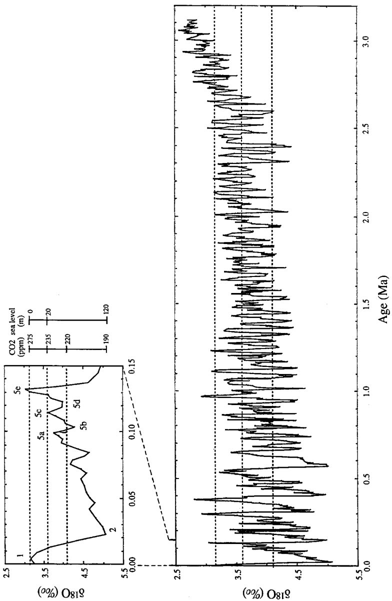

Figure 1 Benthic $\delta^{18}O$ record from DSDP Site 607 (41°N, 33°W) plotted to paleomagnetic timescale of Berggren et al (1985). Dashed horizontal lines (from top to bottom) represent Holocene, stage 5c, and the stage 2/1 boundary at this site. The last climate cycle is expanded to show estimated sea level change and atmospheric CO_2 change. For additional discussion of these data see Raymo (1992).

that prior to 3.1 Ma little evidence for glaciation was observed; that between 2.4 and 3.1 Ma, evidence for small ice sheets equivalent to approximately 40 m of sea level change was observed; and that subsequent to 2.4 Ma, isotopic excursions approximately two thirds of late Pleistocene values were observed, marking "a major change in the character of glaciations." While an apparent age correlation to ice-rafted detritus (IRD) deposits in the North Alantic (Berggren 1972) suggested that the $\delta^{18}O$ variations in V28-179 reflected northern hemisphere climate history, Shackleton & Opdyke were unable to rule out the possibility that the $\delta^{18}O$ record was reflecting ice volume changes in Antarctica.

This opportunity came with the recovery of Site 552, a DSDP (Deep Sea Drilling Project) hydraulic piston core from the North Atlantic, which had relatively higher sedimentation rates and also extended back to the Gauss magnetic chron. At this site, Shackleton et al (1984) were able to show that negative excursions in $\delta^{18}O$ correlated with influx of IRD—unequivocal evidence for nearby continental ice sheets. With this new record, Shackleton et al (1984) reaffirmed their age estimate of $\sim$2.4 Ma for the initiation of major northern hemisphere glaciation. However, the lack of any significant IRD prior to 2.5 Ma led them to conclude that northern glaciation was minimal prior to this time although the $\delta^{18}O$ record suggested that "there was considerable climatic variability, somewhere on the globe, even before 2.5 Myr."

The most recent step forward in our understanding of the evolution of the northern hemisphere ice ages, with respect to the structure of the $\delta^{18}O$ record, was again associated with an improvement in core recovery techniques. Ruddiman et al (1986a) demonstrated that significant amounts of core material could be lost during the hydraulic piston coring process due to ship heaving and other factors. To circumvent this problem they double-cored each site, correlated between holes, and then used material from the offset hole to patch in missing sections from the main hole. Using offset cores, Raymo et al (1989) generated a late Pliocene composite section from Site 607 in the North Atlantic. The more complete $\delta^{18}O$ (Figure 1) and %carbonate records from this site suggested that hiatuses were indeed present at each of the core breaks in the Site 552 record, including a 130 Kyr break from about 2.52 to 2.65 Ma. The presence of this hiatus at Site 552 made the initiation of northern hemisphere glaciation appear more abrupt than it actually was.

The Site 607 record clearly shows "Milankovitch" climate oscillations which get progressively "colder" (i.e. more positive in $\delta^{18}O$) during the late Pliocene (Figures 1 and 2). Between 3.1 and 2.95 Ma, $\delta^{18}O$ values oscillated between minimum values around 2.6‰ and maximum values around 3.1‰. Thus even the cold extremes during this interval were more negative in $\delta^{18}O$ (warmer) than Holocene values ($\sim$3.2‰). Within this interval, data indicate that bottom waters were up to 3.5°C warmer than present or that there was significantly less ice on Antarctica (or some combination of these two effects). Complete deglaciation of the modern Antarctic ice cap would decrease the mean ocean $\delta^{18}O$ value by $\sim$0.9‰ relative to today (Shackleton & Kennett 1975). At 2.95 Ma, slight but obvious steps in both the mean and amplitude of the $\delta^{18}O$ signal occurred at Site 607, and between 2.95 and $\sim$2.7 Ma, minimum $\delta^{18}O$ values were slightly more positive ($\sim$2.8‰) while maximum values hovered around 3.4‰. Thus, it is not until after 2.95 Ma that "cold" episodes were colder than today. Warm intervals remained significantly warmer than the Holocene.

Between 2.7 and 2.4 Ma, small amounts of IRD are observed coincident with positive $\delta^{18}O$ excursions in North Atlantic cores (e.g. 607, 609, and 610) providing direct evidence for continental ice sheets in the northern hemisphere and melting icebergs in the North Atlantic Ocean (Raymo et al 1989, Jansen et al 1990). Large-scale IRD fluxes also appeared in the Norwegian and Barents Seas at this time documenting expansion of northern European ice sheets (Vorren et al 1988, Jansen et al 1988, Jansen & Sjøholm 1991). Interglacial $\delta^{18}O$ values averaged about 3.1‰ at Site 607 while, at the cold extreme, values were $\sim$4.0‰; the overall amplitude was approximately half of the late Pleistocene $\delta^{18}O$ signal. During glaciations between 2.7 and 2.4 Ma, high-latitude climate is inferred to have been significantly colder than at present. However, $\delta^{18}O$ values still fell within the range of oxygen isotopic stage 5 indicating that although continental ice sheets expanded on land, these cold events were more analogous to substages 5b and 5d than to glacial stages 2–4 (Figure 1).

Subsequent to 2.4 Ma, three $\delta^{18}O$ events (stages 96, 98, and 100) with a typical amplitude of $\sim$1.2‰ are correlated with the first major influxes of IRD into the open North Atlantic Ocean. These events, traditionally thought to mark the "onset" of significant northern hemisphere glaciation, reach approximately two thirds of late Pleistocene glacial values (as originally proposed by Shackleton & Opdyke 1977). Minimum values around 3.2‰ indicate that warm periods at this time were very similar to present (with Antarctic ice volumes similar to today). Lastly, the interval between 2.3 and 2.0 Ma was characterized by lower $\delta^{18}O$ amplitudes ($\sim$0.6‰) suggesting that glacial episodes were not as pronounced as between 2.3 and 2.4 Ma (Raymo et al 1989).

All published high resolution $\delta^{18}O$ records (Figures 2 and 3; Shackleton & Hall 1989, Sikes et al 1991, Raymo et al 1992) confirm the basic ice volume history outlined above. The gradual increase in $\delta^{18}O$ values between 3.0 and 2.4 Ma is associated with a gradual increase in IRD in

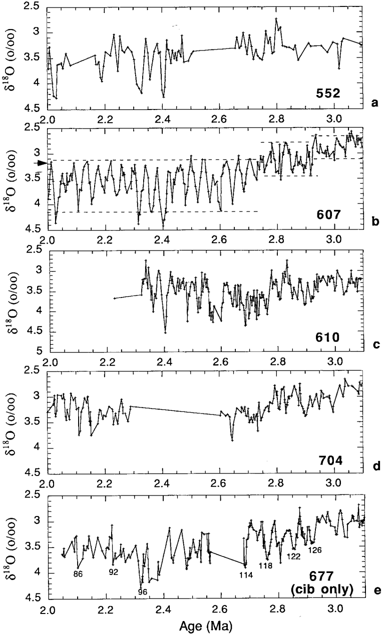

Figure 2 Late Pliocene $\delta^{18}O$ data plotted vs age (see Raymo et al 1992). Gaps represent material inferred to be missing at core breaks. Note that only data from *Cibicidoides* are used in this figure. Arrow indicates mean Holocene $\delta^{18}O$ values.

the Norwegian Sea and the open North Atlantic. While minor amounts of IRD are observed in the Norwegian Sea as early as 5.45 Ma (Jansen et al 1990), it is not till 2.57 Ma (2.44 Ma in open North Atlantic) that the first large fluxes, reflecting widespread continental glaciation, are observed in the Norwegian Sea. Likewise, in the northern Pacific Ocean, the onset of significant ice-rafting episodes is dated at ~2.48 Ma (Rea & Schrader 1985). The de facto onset of northern hemisphere glaciation can be considered to be stage 100, or 2.44 Ma (2.57 Ma by the time scale of Shackleton et al 1990).

The terrestrial record of continental glaciation in the northern hemisphere is obviously less complete and more poorly dated. Early efforts in dating till in Iceland suggested that glacial activity started as early as 3.2 Ma (McDougall & Wensink 1966; see also discussion in Raymo et al 1986). A number of younger tills, dating between 1.7 and 2.5 Ma have been found in the Pacific Northwest and Yellowstone regions (Thompson 1991). In a comprehensive review of vegetation records from the American west, Thompson (1991) documents a trend from generally warmer and wetter conditions between 4.8 and 2.4 Ma to colder, more arid conditions after 2.4 Ma. He also discusses evidence for the transient development of steppe-like environments in Idaho as early as 3.0 Ma. Freshwater ostracode data from the American west also suggest wetter environments between 3.5 and 2.5 Ma (Forester 1991). In north central China, the first loess deposits, formed under extremely cold, dry conditions, are dated at ~2.35 Ma (Kukla 1987, Kukla & An 1989). The Chinese loess is slightly younger than the estimated age of 2.6 Ma for increased dust fluxes to the Pacific Ocean from this region (Rea & Schrader 1985)—a disagreement that may indicate uncertainty in the land-ocean age correlation.

Eolian loess deposits, typically associated with glacial conditions, have also been identified in sediments as old as 3.0 Ma in central Alaska (Westgate et al 1990). These deposits may have been derived from mountain glaciers in the Alaska Range. However, in agreement with oxygen isotope evidence, fossil vegetation from the circum-Arctic shows that the climate was, at least intermittently, significantly warmer than at present in the late Pliocene. A pollen-based climate reconstruction from the Kap Kobenhavn formation of northern Greenland (Funder et al 1985) suggests that conditions similar to those in modern Labrador existed at this time and that the Greenland ice sheet was either absent or very reduced in size. Likewise, sedimentologic structures along the coast suggest at least seasonally ice-free conditions at this time. Although the dating of this section is uncertain and could range fron 3.1 to 1.6 m.y., an age of about 2.0 Ma BP has been proposed for this site (Funder et al 1985, Brigham-Grette & Carter 1992). At Ocean Point in northern Alaska, pollen recon-

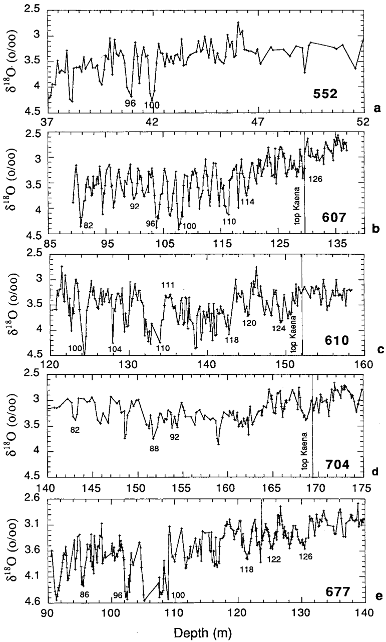

Figure 3 Late Pliocene $\delta^{18}O$ records from Sites 552, 610, 607, 704, and 677 plotted against actual sub-bottom depth (552, 610) or to composite sub-bottom depth (607, 704, 677). The top of the Kaena magnetic subchron (2.92 Ma) is indicated in those cores with a magnetic record. Selected isotopic stages are identified according to Raymo et al (1989, 1992).

structions from the Gubik Formation also suggest a warmer climate, represented by an open boreal forest, in the late Pliocene (Nelson & Carter 1985, Brigham-Grette & Carter 1992). The absolute age of this section is again not well-constrained and is placed sometime between 2.7 Ma and 2.1 Ma (Brigham-Grette & Carter 1992). During warm intervals it is unlikely that seasonal Arctic sea ice existed at this coastal location.

In the Fish Creek section, also in northern Alaska, invertebrate (mollusc) and vertebrate (sea otter) evidence, shown to be younger than 2.5 Ma by magnetostratigraphy, indicate that the Arctic margin could not have been frozen for more than one month a year at this location. Similarly, geologic evidence from a number of other sites, summarized by Repenning et al (1987) and Brigham-Grette & Carter (1992) reflects a generally warmer circum-Arctic climate in the late Pliocene, with some indications by pollen of alternating warmer and colder intervals. Unfortunately, the ages of most of these sites are poorly constrained but are believed to have been deposited prior to the Pliocene-Pleistocene boundary (1.6 Ma). Because no land sections have been firmly dated in the critical late Pliocene/early Pleistocene interval after 2.5 Ma, it remains uncertain when the present-day perennial sea ice cover developed in the Arctic. The deep sea record from this ocean is also ambiguous, primarily due to low sedimentation rates and difficulties in accurately dating these sediments. Herman & Hopkins (1980), Herman et al (1989), and Gilbert & Clark (1982/1983), all proposed that modern Arctic sea ice cover formed after 0.9 Ma ago. However, permanent perennial sea ice may have developed as early as the Pliocene/Pleistocene boundary (1.6 Ma; Scott et al 1989). It appears that the early period of northern hemisphere glaciation may have been associated with only seasonal Arctic sea ice cover, at least during Milankovitch interglacial extremes.

RESPONSE OF GLOBAL CLIMATE TO GLACIATION

Sea Surface Temperatures

A number of studies have examined the response of sea surface temperatures (SSTs) to northern hemisphere glaciation. In addition, climate modelers have recently expressed an interest in the response of ocean temperatures to the warm climates of 3.0 Ma ago, warmth possibly caused by enhanced atmospheric CO_2 levels (e.g. Crowley 1991). Studies of foraminiferal assemblages from the late Pliocene (3.4–1.6 Ma; Berggren 1972, Poore & Berggren 1975, Loubere & Moss 1986, Raymo et al 1986) suggest that North Atlantic sea surface temperatures were warmer prior to 2.5 Ma. This inference was based primarily on the fact that larger, more ornate species (e.g. *N. atlantica*) were replaced at high latitudes by smaller, denser, more compact species (e.g. *N. pachyderma* sinistral) as ice sheets developed in the late Pliocene. Unfortunately, because many of the most abundant late Pliocene species (*N. atlantica*, *G. puncticulata*, *N. acostoensis*, etc) became extinct by the Pleistocene (1.6 Ma), a direct calibration of the temperature and salinity preferences of these species is impossible. With this in mind, I discuss below recent studies that reconstruct sea surface temperatures for the mid-Pliocene North Atlantic. In particular, as climate modelers are now using these reconstructions as input into general circulation model experiments, an assessment of the assumptions these studies are based on seems warranted.

Many readers are probably familiar with the CLIMAP (1981) global SST reconstruction for the last glacial maximum (18,000 years ago). The CLIMAP group used transfer functions developed by Imbrie & Kipp (1971) and Kipp (1976) to infer past sea surface temperatures by looking at the geographic distribution in the past of modern species whose environmental tolerances are well-known. Specifically, in Kipp (1976), the percent abundance of 29 foraminiferal species was estimated in 191 core-top (modern) samples from the North Atlantic. These data were factor-analyzed into varimax assemblages and a least-squares regression technique was used to relate the varimax assemblages to observed sea surface temperatures and salinities at each core site. From these relationships, transfer functions were produced. Downcore species abundance data are then described in terms of the core-top assemblages, which are then used with the paleoecological equations to estimate past environmental conditions at a site. The standard errors of the transfer function were estimated by Kipp (1976) to be ± 1.16°C for the cold season and ± 1.38°C for the warm. On rare occasions, one can find assemblages in late Pleistocene sediments that have no modern analog (due to dissolution, stratigraphic mixing, taxonomic misidentification, local ecological variation, or other factors). These samples are easily identified by their low communalities—a statistical measure of how well the assemblage fits the varimax model. Imbrie et al (1973) suggest that communalities greater than 0.8 are necessary for accurate environmental estimates.

Of course, the key assumption in this and other transfer functions that have been developed to study Pleistocene climates is that species do not change their environmental preferences with time. If we know that left-coiling *N. pachyderma* has a maximum abundance in water at less than 5°C today, then we can assume that the same holds true for *N. pachyderma* sinistral which lived one million years ago. Difficulties arise when components of the faunal assemblage are extinct since we cannot be assured that they have the same environmental preferences as their extant relatives.

Dowsett & Poore (1990) and Dowsett (1991) developed and tested a foraminiferal transfer function that would be applicable to the Pliocene. The standard errors for their GSF18 transfer function were estimated to be ± 1.47°C for the cold season and ± 1.36°C for the warm season, only slightly different from those of the Kipp (1976) transfer function. To develop GSF18, Dowsett made two important changes from the CLIMAP transfer function. First, he simplified the taxonomy slightly by lumping certain abundant species, such as right and left-coiling *N. pachyderma*, thus reducing the number of taxonomic catagories that needed to be counted. [Ruddiman & Esmay (1986) carried out a similar exercise developing a five "species" transfer function that could be applied much more rapidly to the long Pleistocene sections recovered by DSDP and ODP (Ocean Drilling Program) drilling.] Dowsett (1991) correctly tested the effects of this change in the structure of the transfer function by taking existing faunal data from the 18 Kyr and 122 Kyr time-slice data bases and regrouping them according to his taxonomic scheme. He found, as expected, some loss of resolution in the SST estimates. In particular, lumping cold *Neogloboquadrina* species resulted in an overestimate of high-latitude SSTs by 1–2°C and an underestimate of SSTs at mid-latitudes (30–40°N) by 1–3°C over CLIMAP estimates.

While the above lumping of certain species inevitably leads to a loss of information, this is not the major source of error in GSF18. To generate a Pliocene transfer function, one other assumption is made by Dowsett & Poore (1990), namely that extinct species occupy the same environmental niche as their presumed modern descendents (lineages are inferred from similar gross morphology and biogeographic ranges). This assumption obviously begs the question of why species evolved. Two rationales are offered in support of this assumption: similar morphology and similar geographic ranges. However, while *G. ruber* and the extinct *G. obliquus* have a similar morphology and are assumed to have the same environmental tolerances, a study of their evolutionary lineages (Kennett & Srinivasan 1983) show them to be completely separate over the Neogene (the last 24 Ma). How likely is it that they occupied the same enviromental niche in the Pliocene (*G. obliquus*) and Pleistocene (*G. ruber*)? In most cases, Pliocene species are larger, more delicate, or more ornate than their presumed Pleistocene counterparts (e.g. *G. puncticulata*—*G. inflata*, *N. atlantica*—*N. pachyderma*, *G. miocenica*—*G. menardii*, *G. fistulosis*—*G. sacculifer*) suggesting, by analogy with the modern equator to pole changes in morphology, that the extinct species lived in warmer water. The second rationale used for assuming similar temperature preferences—similar biogeographic ranges—is clearly circular reasoning. How would one recognize a regional water mass change if the de facto assumption is made that extinct species which lived in a region had the same temperature and salinity preferences as the modern species that live there?

How pervasive are now-extinct species in mid-Pliocene planktonic foraminiferal assemblages? At Site 552, one of the sites for which faunal census data were published (Dowsett & Poore 1990), extinct species comprise over 50% of the total fauna in the interval between 2.85 and 3.15 Ma. However, because the transfer function artificially treats these species as their modern-day counterparts, the samples all have communalities above 0.7. At many other sites used in the Dowsett et al (1992) reconstruction (e.g. 502, 548, 606) communalities below 0.7 are quite common (38–56% of samples; Dowsett & Poore 1991), due primarily to "no-analog" percent abundances of the assumed "modern" species. These data, obviously a significant fraction of the Pliocene faunal record, are excluded from the SST reconstruction, possibly imparting a systematic bias to the results.

Taking GFS18 results at face value, and not assigning any error to the substitution of extinct for modern species, Dowsett et al (1992) presented a mid-Pliocene sea surface temperature reconstruction that showed SSTs at 50°N approximately 4°C warmer and SSTs at low latitudes approximately 1°C warmer than at present (Figure 2 in Dowsett et al 1992). Considering the above assumptions, the true estimation error on the GFS18 transfer function is unclear. For instance, the species-lumping discussed above may impart a 1–2°C overestimatation of SSTs at the higher latitudes; likewise, many of the extinct tropical species of the late Pliocene (*G. obliquus*, *G. altispira*, *G. trilobus*, *N. humerosa*, *G. pseudopima*, *G. decoraperta*, *G. miocenica*, *S. seminulina*, *G. limbata*, etc) may actually have preferred warmer water than their Pleistocene replacements. Given the importance of accurate SST reconstructions for the evaluation of climate change mechanisms (e.g. Rind & Chandler 1991, Crowley 1991; see also below), it is critical that vigorous investigation into the preferred habitats of extinct species begin. In particular, estimates of past CO_2 levels using the $\delta^{13}C$ of marine organic matter (e.g. Rau et al 1991, Freeman & Hayes 1992) require accurate estimates of SST to calculate CO_2 solubility in past oceans.

North Atlantic sea surface temperatures have also been estimated using marine ostracod fauna (Cronin & Dowsett 1990, Cronin 1991), which live in shallow water above the seasonal thermocline. In other words, ostracod live at the same depths as planktonic foraminifera, only in a near-shore environment. Using a transfer function based on 100 modern sediment samples divided into 59 ostracode taxa, Cronin (1991) examined Pliocene sections from the tropics (9°N) to polar regions (66°N) along the western Atlantic margin. Based on the distribution of genera at eight well-dated sites, Cronin (1991) concluded that, during the Middle Pliocene (3.5–3.0 Ma), tropical and subtropical shelf-water temperatures were slightly cooler than today, while at temperate to subpolar latitudes temperatures were up to 10°C warmer than today. Temperatures off northern Iceland were up to 6°C warmer than today in summer (4°C in winter) suggesting very little influence by the East Greenland Current or seasonal sea ice which dominates this region today.

Unlike foraminiferal transfer functions, which are based primarily on species, the ostracod transfer function analyzes the fauna at a generic level. By breaking the fauna down into 59 primarily generic groupings, this approach may be more robust with respect to extinct species than the foraminiferal transfer function. Unlike some foraminiferal genera (e.g. the Neogloboquadrinids), ostracode genera are generally restricted to one or only a few climatic zones, possibly resulting in less of a systematic error due to species-level evolution. Cronin (1991) gives an error on the transfer function temperature estimates of ± 2°C.

A study of dynocyst abundances in the northern Atlantic also suggests warmer SSTs prior to 2.4 Ma, although Edwards et al (1991) discuss the pitfalls of inferring environmental conditions from "no-analog" fauna. Based on the presence of subtropical taxa at Site 552 (56°N, 23°W), they suggest a greater northward extension of the Gulf Stream/North Atlantic Current system around 3.0 Ma. Edwards and coworkers also propose that outflow of low-salinity polar water from the Greenland Sea was greatly reduced at this time.

The various methods used to estimate SSTs all generally conclude that the North Atlantic was warmer in the mid-Pliocene around 3.0 Ma ago. This warmth is attributed to enhanced heat transport in the Gulf Stream/North Atlantic Drift system and a weakening of the cold East Greenland Current (Cronin 1991, Edwards et al 1991, Dowsett et al 1992). A weakening, or warming, of the East Greenland Current is also supported by evidence for Arctic coastal warmth discussed in the previous section. Although warmer SSTs could also be ascribed to decreased heat removal by the atmosphere (especially in winter), further evidence suggesting stronger ocean heat transport to the North Atlantic is discussed in the following section on thermohaline circulation. In detail, the ostracod transfer function suggests colder low-latitude regions (although cooler termperatures could be due to stronger summer upwelling along the coast) and much warmer mid- to high-latitude SSTs than the foraminiferal-based estimates. At present, it is impossible to say which estimates are more accurate.

In the North Pacific, the SST response to glacial inception in the late Pliocene has been investigated using siliceous flora (diatoms) and fauna (radiolaria) (Morley & Dworetzky 1991). Based on assemblage abundances, these authors suggest a pattern of climate change very similar to that inferred from North Atlantic fauna as well as oxygen isotopes: a period of relatively warmer, mild conditions between 3.0 and 2.7 Ma, followed by a major cooling at 2.46 Ma, roughly coincident with the major enrichment in benthic oxygen isotope records discussed above as well as with large inputs of IRD into both the North Atlantic (Shackleton et al 1984, Raymo et al 1989) and Pacific Oceans (Rea & Schrader 1985). As in the North Atlantic $\delta^{18}O$ and IRD records, these authors note a brief return to milder, warmer conditions between $\sim$2.1 and 2.3 Ma.

Thermohaline Circulation

Ocean thermohaline circulation plays an important role in controlling the global distribution of heat and in modulating exchange of CO_2 between the deep ocean and the atmosphere (e.g. Broecker & Denton 1989, Rind & Chandler 1991, Boyle 1988, Broecker & Peng 1989, Charles & Fairbanks 1992). Today, most deep water forms in two locations: the North Atlantic Ocean and the Southern Ocean around Antarctica. The sinking of deep water in the Norwegian-Greenland Seas and Labrador Sea sets up a "conveyor belt" which then draws additional warm, salty thermocline water northward. As this water cools and sinks, heat released to the atmosphere is advected eastward over Europe and Scandanavia warming these regions. During the last glacial maximum, North Atlantic Deep Water (NADW) formation was greatly suppressed (Curry & Lohmann 1983; Boyle & Keigwin 1982, 1987; Oppo & Fairbanks 1987) and sea surface temperatures dropped as the polar front, or sea ice limit, migrated southward to the latitude of Spain (CLIMAP Project Members 1981). The decline in NADW formation may also have led to the reduction of atmospheric CO_2 observed in ice cores, thereby causing further cooling (Boyle 1988, Keir 1988, Broecker & Peng 1989).

Because ocean thermohaline circulation affects both the global surface heat distribution as well as atmospheric CO_2 concentrations, the response of NADW formation to global "greenhouse" warming is an important question. Modern oceanographic studies by Brewer et al (1983), Aagaard (1988), and Schlosser et al (1991) suggest that the salinity of water in the North Atlantic plays a vital role in controlling thermohaline overturn. For instance, a low-salinity anomaly in the North Atlantic, resulting from increased freshwater flow from the Arctic Ocean, was associated with a reduction in deep-water formation in the 1970s. Likewise, an ocean general circulation model suggested that thermohaline overturn would decrease in the North Atlantic in a doubled CO_2 world (Mikolajewicz et al 1990). This change was attributed to a decrease in surface water salinity caused by increased regional precipitation. Because weaker NADW production could decrease the transport of heat to high latitudes as well as lower atmospheric CO_2 (as observed during last glacial maximum), the simplest interpretation of these model results is that weaker thermohaline circulation could act as a negative feedback as the climate warmed. However, the geologic record suggests the opposite, warmer climates are associated with stronger NADW formation, at least for the last 3.2 Ma.

The history of glacial-interglacial change in deep ocean circulation is reconstructed with the use of carbon isotopes recorded in the calcite tests of foraminifera that live on the ocean floor (see Curry et al 1988 for an extensive review). Because NADW forms with high initial $\delta^{13}C$ values ($>1.0‰$), its presence is clearly seen in vertical cross sections of $\delta^{13}C$ in the modern ocean (Kroopnick 1985). Upper Circumpolar Deep Water (UCDW) and Antarctic Bottom Water (AABW), the other two major deep-water masses in the Atlantic, are more negative in $\delta^{13}C$. By reconstructing deep-water $\delta^{13}C$ gradients in the past, paleoceanographers can determine how the path of deep-water flow and the relative contribution of different source waters to the deep ocean changed through time.

By examining the evolution of $\delta^{13}C$ gradients between three cores from the North Atlantic and Pacific oceans, Raymo et al (1990a) reconstructed the history of NADW formation back to 2.5 Ma. They showed that NADW formation typically decreased during glaciations and that the suppression of NADW became particularly pronounced over the last million years with the development of larger ice sheets. Raymo et al (1992) studied the response of NADW to global cooling and the intensification of northern hemisphere glaciation between 3.1 and 2.0 Ma. They showed that global cooling led to gradually stronger suppression of NADW production and that the North Atlantic, prior to major northern hemisphere glaciation, was probably characterized by vigorous thermohaline circulation. (For a discussion of compatible and conflicting sedimentological studies see that paper.) This interpretation, also reached by Hodell & Venz (1992), rests heavily on the carbon isotopic record of ODP Site 704 from the South Atlantic (Figure 4). Unfortunately, this record has a significant hiatus around the late Gauss/early Matuyama and, as a result, stratigraphic control is poor. In addition, Mix et al (1994) suggest that Site 704 may be sensitive to regional variations in Antarctic circumpolar circulation which mask the true flux of NADW from the North Atlantic. Work in progress on ODP Leg 108 cores at the equator should provide additional constraints on the response of NADW formation to northern hemisphere glaciation.

What caused the inferred decrease in NADW formation over the late Pliocene? By analogy to the late Pleistocene study of Boyle & Keigwin (1987), Raymo et al (1992) proposed that lowered sea surface temperatures

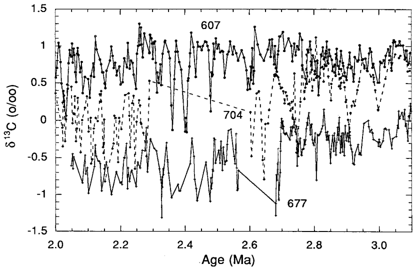

Figure 4 *Cibicidoides* $\delta^{13}C$ data from Sites 607 (North Atlantic), 704 (South Atlantic), and 677 (Equatorial Pacific). Site 704 $\delta^{13}C$ values become increasingly Pacific-like after 2.75 Ma suggesting decreased production of NADW.

reduced evaporation and, hence, decreased the surface salinity and potential density of surface waters. These waters were therefore less likely to sink and form a deep-water mass. As discussed above, prior to 2.6 Ma, planktonic fauna and ostracod data indicate warmer SSTs in the North Atlantic (Raymo et al 1986; Loubere & Moss 1986; Dowsett & Poore 1990, 1991; Cronin & Dowsett 1990; Cronin 1991), while many other high-latitude sites suggest that the Arctic was seasonally (or completely) ice free at this time (Funder et al 1985, Brouwers et al 1991, Matthews & Ovenden 1990, Carter et al 1986). Subsequent cooling is observed at many of these locations (North Atlantic and Arctic) in conjunction with the intensification of northern hemisphere glaciation around 2.4 Ma.

Two mechanisms that could decrease SST and deep-water production rates have been suggested by general circulation model (GCM) results. Manabe & Broccoli (1985) showed that orographic diversion of winds by a continental ice sheet on Canada would result in a significant cooling of the surface North Atlantic. Likewise, a GCM experiment in which Arctic sea ice extent was reduced (Raymo et al 1990b) (reflecting climate conditions in the pre-glacial Pliocene) showed three changes that would pro-

mote greater thermohaline overturn in the Norwegian-Greenland and Labrador Seas and, thus, more vigorous formation of NADW: (*a*) an enhancement of surface water salinities in the North Atlantic region resulting from an increase in evaporation relative to precipitation; (*b*) a localized strengthening of the Icelandic low over the Norwegian-Greenland Sea which would enhance northward advection of salty water into the area of deep-water formation; and (*c*) an increase in the salinity of water leaving the Arctic, driven by increased evaporative fluxes in this region when sea ice limits are reduced. Further studies, both of the paleoceanographic record and with coupled ocean-atmosphere climate models, are needed to evaluate the above possibilities.

Although here we consider changes in thermohaline circulation to be a response to glaciation, they can just as easily be considered as a potential cause for glaciation (see later in this paper). A decrease in the production of NADW would decrease heat transport to high latitudes as well as possibly result in a decrease in atmospheric CO_2—both factors that would enhance cooling. Thus, at the least, NADW production can act as a strong positive feedback and possibly as an independent climate forcing mechanism (e.g. Rind & Chandler 1991).

Antarctic Glacial History

As discussed earlier, benthic $\delta^{18}O$ records from the mid-Pliocene are consistent with a major deglaciation of Antarctica around 3.0 Ma (although more negative $\delta^{18}O$ values at this time could also be ascribed to warmer deep ocean temperatures). However, estimates that eustatic sea level was up to 40 m higher in the Middle Pliocene also point to significant deglaciation of Antarctica at this time (Haq et al 1987, Dowsett & Cronin 1990, Krantz 1991). A pronounced transgression, which deposited the Rushmere and Morgarts Beach members of the Yorktown Formation, occurred between 4.0 and 3.2 Ma. Krantz (1991) estimates sea level up to 35 meters higher than present during this interval. Studies of the Orangeburg Scarp in North and South Carolina also suggest sea levels up to 35 meters higher than present between 3.5 and 3.0 Ma (Dowsett & Cronin 1990).

More direct evidence for Antarctic deglaciation comes from terrestrial plant remains which were deposited in the Transantarctic Mountains after 3.0 Ma ago (Webb et al 1987). These deposits contain *Nothofagus*, or southern beech, a plant which suggests a climate more analogous to that of southern Patagonia. Biostratigraphic diatom markers (Webb et al 1984) and radiometric dating of coexisting ash layers in Antarctica (Barrett et al 1992) both confirm that the warmer conditions existed later than three million years ago. Such a major deglaciation of East Antarctica (possibly more than half the current ice sheet volume), occurring when global climate was only slightly warmer than at present, implies a cryosphere much less stable than previously thought (Kennett 1977). In particular, Barrett et al (1992) caution that ice sheet stability needs to be evaluated more seriously in face of the greenhouse warming predicted for the next century. They propose that the Antarctic ice sheet reached its present size simultaneously with the development of large ice sheets in the northern hemisphere ($\sim$2.4 Ma).

The evidence for a major Pliocene deglaciation of Antarctica is not universally accepted. Geomorphologic features and glacial landforms show evidence for continuous cold environments for at least the last 10 Ma with no suggestion of meltwater activity or temperate climate conditions (Denton et al 1984, Clapperton & Sugden 1990). Likewise, a number of paleoclimatologists argue that since Antarctica is currently the driest continent on Earth with mean annual temperatures well below zero, any warming would actually increase ice volume by increasing snowfall (e.g. Prentice & Matthews 1991, Oglesby 1989). This line of reasoning would predict that Antarctic ice volume decreased over the late Pliocene interval of polar cooling ($\sim$3.0–2.4 Ma).

In contrast, studies of subantarctic deep-sea sediments (Hodell & Ciesielski 1990, 1991; Hodell & Venz,1992; Warnke & Allen 1991) indicate increased delivery of ice-rafted sediment after 2.46 Ma, suggesting ice sheet expansion on Antarctica. This followed an interval of progressive cooling from 3.25 Ma to 2.4 Ma over which the Polar Front Zone migrated northward in the Atlantic sector of the Southern Ocean and silica productivity increased (Hodell & Warnke 1991). Froelich et al (1991) inferred from high rates of carbonate deposition that Site 704, located just north of the polar front today, was also north of the polar front between 2.9 and 2.7 Ma (e.g. conditions similar to today). Perennial sea ice cover was also established in the Weddell Sea at this time (Abelmann et al 1990). Carbon isotopic data indicate decreased ventilation of deep waters in the circum-Antarctic after 2.75 Ma, undoubtedly due to a reduction in North Atlantic Deep Water flux during glaciations (Hodell & Ciesielski 1990, 1991; Hodell & Venz 1992; Raymo et al 1992). Hodell & Ceisielski (1990) propose that the climate of the southern polar regions is linked to that of the northern hemisphere through a series of positive feedbacks. In their model, increased suppression of NADW during northern hemisphere glaciations decreases the flux of salt and heat to the Southern Ocean allowing the northward expansion of sea ice. In addition, they point out that a decrease in the flux of NADW could result in lowered atmospheric CO_2 [via the polar alkalinity model of Broecker & Peng (1989)] further strengthening the climate coupling between the northern and southern hemispheres.

Low Latitude Climate

One of the most pronounced features of low to mid-latitude climate is the seasonal change in winds and precipitation associated with the Asian monsoon. The monsoon, caused by the differential heating of land (primarily the Tibetan Plateau) and sea (the Indian Ocean), is characterized by rising air masses over land in summer and moisture-laden surface winds that flow off the Indian Ocean into southern Asia. These southwesterly winds bring the summer monsoon rains. In winter, as land surfaces cool, the seasonal wind pattern reverses. It has been shown that throughout Africa, the Mediterranean, and Asia, regional precipitation and aridity variations are tied to the strength of the Asian monsoon (Street-Perrott & Harrison 1984, Pokras & Mix 1985, Prell & Van Campo 1986, Clemens & Prell 1990). In addition, the intensity of oceanic upwelling in the Arabian Sea also reflects the strength of the summer monsoon winds (Prell 1984, Prell & Kutzbach 1987).

The strength of the monsoon, in turn, has been shown to vary as a function of three primary factors: the strength of seasonal insolation over the Tibetan plateau, plateau elevation, and the presence of glacial or interglacial surface boundary conditions (see review by Prell & Kutzbach 1992). Using GCM experiments, Kutzbach (1981) showed that stronger insolation heating at the latitude of the plateau, induced by Milankovitch variations in precession, resulted in a stronger monsoon—a prediction matched by observations. Likewise, a series of GCM experiments that varied Tibetan Plateau elevation (Hahn & Manabe 1975, Ruddiman & Kutzbach 1989, Prell & Kutzbach 1992) suggest that monsoon intensity increases with the elevation of the plateau. Lastly, the presence of continental ice sheets in the northern hemisphere acts to suppress the monsoon as indicated by evidence for a weaker monsoon during the Last Glacial Maximum, a time of similar low-latitude insolation (Prell 1984, Street-Perrott & Harrison 1984, Prell & Van Campo 1986). This suppression may be due to glacial changes in SST, atmospheric CO_2, land albedo, and/or the extent of seasonal snow in Eurasia and Tibet rather than the extent and height of the Laurentide or Scandanavian ice sheets per se (Prell & Kutzbach 1992).

Studies of the long-term evolution of the Asian monsoon indicate that the onset of northern hemisphere glaciation resulted in increased aridity in northeast Africa after 2.4 Ma, possibly due to glacial suppression of monsoon intensity. This trend is seen in palynological studies (Bonnefille 1983), terrestrial isotopic studies (Cerling et al 1977, Abell 1982), and macrofaunal studies (Vrba 1985, Wesselman 1985, Grine 1986). In the eastern equatorial Atlantic, a pronounced increase in the input of terrigenous dust is observed at 2.4 Ma—an increase associated with strengthened northern hemisphere winter trade winds (Ruddiman & Janecek 1989). Enhanced opal fluxes suggest that upwelling intensity also increased after 2.5 Ma, again probably due to strengthened trade winds (Ruddiman & Janecek 1989). After 2.4 Ma, the modulation of dust fluxes to the Arabian Sea at the 41,000 year obliquity period of ice sheet variablility also points to high-latitude influence on African climate and monsoonal variability after this time (de Menocal et al 1991). de Menocal et al (1993) propose that low-latitude African climate was dominated by monsoon forcing, at precessional frequencies, prior to the initiation of major northern hemisphere glaciation and that, after that time, regional precipitation and aridity patterns at low latitudes were more strongly influenced by remote forcing from high latitudes at the obliquity frequency. Additional high-resolution records of Pliocene low-latitude climate are needed to test this hypothesis.

CAUSE OF NORTHERN HEMISPHERE GLACIATION

As outlined above, northern hemisphere glaciation began gradually, rather than abruptly, between 2.9 and 2.4 Ma. Mechanisms to explain the onset of the ice ages fall within two broad catagories: terrestrial and extraterrestrial. Proposed extraterrestrial causes of climate change include variations in solar output, collisions with asteroids, passage through interstellar dust clouds, as well as supernova explosions, to name a few (see Pollack 1982 for an excellent summary). However, not one of these factors has provided a convincing explanation for the onset of northern hemisphere glaciation. While Milankovitch variations in insolation caused by changes in the Earth's obliquity, precession, and eccentricity, obviously play a critical role in pacing the sequence of glacial/interglacial oscillations, they can not explain the long-term cooling trend observed in the Pliocene-Pleistocene. Below, explanations for northern hemisphere glaciation that invoke terrestrial changes in boundary conditions are presented and discussed. In evaluating these mechanisms, keep in mind two possible models of the climate system: one in which climate responds linearly to changes in forcing; and one in which climate "thresholds" are crossed. In the second case some kind of climate instability results in a nonlinear response to external forcing (e.g. North 1984).

Plate Movements

Many of the terrestrial mechanisms proposed as explanations for long-term climate variation are related to plate tectonics and the dynamic forces within the Earth that are continually modifying the Earth's surface.

Probably one the simplest ideas put forth to explain the timing of glaciations is polar positioning of continents. Very few studies of Paleozoic climate history fail to mention this popular idea (e.g. Crowell & Frakes 1970, Caputo & Crowell 1985, Crowley et al 1987), however, as we move toward the present, it becomes less certain that polar continental position has any strong effect on climate. Barron (1981) and Barron & Washington (1985) conclude that while long-term plate motions could be responsible for the general climate cooling observed over the last 150 Ma, such movements would be too slight to account for the relatively faster cooling observed since the Eocene. The only way plate positions can effect a relatively rapid change in climate is by invoking a "critical point" in the climate system. North (1984) and North & Crowley (1985) discuss how small icecap instabilities could result in a nonlinear response to subtle variations in seasonal insolation. They speculate that such an instability could lead to the rapid formation of the Greenland ice sheet. If one then invoked a series of positive feedbacks, such as the development of perennial Arctic sea ice cover or a decrease in NADW production, such a change could conceivably then result in widespread northern hemisphere glaciation.

Sills and Gateways

For many years researchers have speculated that the onset of northern hemisphere glaciation was due to the uplift of the Panamanian Isthmus or the subsidence of the Bering Straits. Age estimates for the final closure of the Panamanian Isthmus range from 2.5 to 3.7 Ma. Based on the appearance of North American large mammal assemblages in South America (and vice-versa), faunal exchange between the two continents happened between 2.5 and 2.8 Ma (Lundelius 1987, Marshall 1988, MacFadden et al 1993). By contrast, marine evidence suggests that uplift of the Isthmus had a significant impact on ocean circulation and biotic exchange much earlier (Duque-Caro 1990, Brunner 1984, Emiliani et al 1972). Based on carbon and oxygen isotopic data and coiling ratios of planktonic foramanifera, Keigwin (1978 1982) estimated that surface waters of the Caribbean and Pacific were isolated by 3 Ma, while a study of shallow-water fauna on both sides of the Isthmus concluded that it was effectively closed by 3.5 Ma (Coates et al 1992). It is unclear how the marine and terrestrial estimates can be reconciled although one can speculate that mammal migrations could not occur until the bridge was emergent everywhere.

The relationship between the development of the Isthmus of Panama and the initiation of northern hemisphere glaciation is also unclear. The formation of the Isthmus could predate major northern hemisphere glaciation by as much as a million years or, if the mammal data are taken at face value, it could have occurred almost simultaneously with global cooling. Early researchers suggested that a closed isthmus would deflect warm Gulf Stream water to the northwest Atlantic providing a ready source of moisture for ice growth (e.g. Stokes 1955). However, an ocean circulation model (Maier-Reimer & Mikolajewicz 1990) suggests the opposite: The North Atlantic region warms significantly when the Isthmus is closed due to stronger ocean heat transport. The absence of a strong meridional current system results in regional cooling and the growth of sea ice in the Norwegian Sea. To explain higher polar temperatures in the early Pliocene, Maier-Reimer & Mikolajewicz conclude that some other factor, such as higher atmospheric CO_2, is needed to provide warmth when the Isthmus was open.

In the northern Pacific Ocean, the opening of the Bering Straits appears to have occurred earlier than the climate cooling which began around 2.9 Ma and culminated in major continental ice growth by 2.4 Ma. Atlantic and Arctic marine species in Middle Pliocene deposits from eastern Kamchatka indicate that faunal migrations were occuring through the Bering Straits by 4.1 Ma (Gladenkov et al 1991). Additional migrations are documented between 3.6 and 3.2 Ma when many Pacific species invade the Arctic and North Atlantic Oceans (Hopkins 1967, Gladenkov 1981). Thus, changes in moisture and heat transport through the Bering Straits are not a likely explanation for late Pliocene cooling of the northern hemisphere.

Topographic Changes

Another important aspect of the Earth's geography is its vertical dimension, e.g. mountains and plateaus. Many early ice age theories focused on mountains, undoubtedly inspired by the presence of majestic alpine glaciers in many of the mid- to high-latitude regions of the world. One such mechanism invokes epierogenic uplift of northern Canada causing nucleation of continental ice sheets and, ultimately, glaciation (Flint 1957, Birchfield et al 1982), although relatively little data exist to support this hypothesis. Ruddiman et al (1986b) and Ruddiman & Raymo (1988) proposed that Pliocene uplift of the Tibetan and Colorado Plateaus could have initiated northern hemisphere glaciation via the effects of uplift on planetary wave structure. By enhancing southward meanders of the upper westerlies and the outbreak of cold polar air masses over North America and Europe, plateau uplift could have led to glacial inception. However, a problem with this hypothesis is that the evidence is weak for recent rapid plateau uplift in Tibet and the American west. Molnar & England (1990) argued that much of the evidence for Pliocene-Pleistocene uplift could be accounted for by enhanced erosion and generation of relief rather than by a dramatic increase in mean elevations. In particular, the Tibetan Plateau probably attained much of its present elevation by the late Miocene (Sorkabi & Stump 1993). More recent work by Ruddiman and colleagues have emphasized the role of plateau uplift in explaining the evolution of regional precipitation and temperature patterns and global climate over the last 40 Ma, rather than the last 3 Ma (Ruddiman & Kutzbach 1989, Ruddiman et al 1989, Prell & Kutzbach 1992).

Volcanism

From the above discussion it would appear that in addition to geography and topography, other factors must play a role in explaining the onset of northern hemisphere glaciation. One possibility is that global cooling resulted from enhanced volcanism and the corresponding increase in ash and aerosol concentrations in the atmosphere which would reflect sunlight. Based on the distribution of ash layers in ocean sediments, Kennett & Thunell (1977) noted a general increase in global volcanism in the Quaternary. However, this and other early surveys (Rea & Scheidegger 1979) suggest that the increase in volcanism occurred after the Pliocene-Pleistocene boundary at 1.6 Ma. Likewise, it was argued that conversion of ash to bentonites and the movement of sea floor crust towards island arc sources of ash would bias the ash records toward a late Neogene increase in volcanism (Ninkovich & Donn 1976, Hein et al 1978). It was also questioned whether volcanism could provide the persistent climate forcing needed to explain millions of years of cold glacial climates or whether it acted as a trigger, tipping the system into a new regime.

The above questions have recently come to the forefront with the presentation of initial drilling results from North Pacific ODP Leg 145. Rea et al (1993) show evidence for a several-fold increase in regional volcanism essentially coincident with the initiation of northern hemisphere glaciation. Undoubtedly the volcanism-climate link will be examined anew as these results are published over the next few years.

Atmospheric Composition

It has long been recognized that variations in radiatively important trace gases such as carbon dioxide or water could have significant effects on the thermal radiation balance of the Earth's atmosphere and, hence, global temperatures (e.g. Chamberlin 1899). This link is supported by the strong covariance of atmospheric CO_2 and temperature observed in the Vostok ice core record of the last 140,000 years (Barnola et al 1987). A number of hypotheses have been proposed that link the longer-term evolution of global climate to changes in the composition of the Earth's atmosphere;

in all cases, these mechanisms have plate tectonic motions as their root cause (e.g. Walker et al 1981, Raymo et al 1988, Berner 1990).

Of particular relevance is the study of Raymo et al (1988) which summarized evidence that global chemical weathering rates had increased significantly over the last 5 Ma. Because chemical weathering is the major sink of atmospheric CO_2 on geologic time scales, they proposed that this increase in weathering led to a "reverse greenhouse" and global cooling since the late Miocene. This idea found support in the study of Crowley (1991) who compared evidence for Middle Pliocene warmth with the results of general circulation model experiments and concluded that the climate of three million years ago was consistent with doubled CO_2 levels (e.g. approximately 550 ppm CO_2 by volume). Recent attempts to test this hypothesis using a proxy method based on the $\delta^{13}C$ of marine organic carbon [see Rau et al (1991) for a description] suggest that mid-Pliocene atmospheric CO_2 levels were closer to $\sim$360 ppm, approximately 25% higher than preindustrial values (Raymo & Rau 1992). These preliminary results await confirmation.

Ocean Heat Transport

An alternative hypothesis to "reverse greenhouse" cooling was proposed by Rind & Chandler (1991). They suggested that pre-Pleistocene warmth at high latitudes was due to stronger ocean thermohaline circulation which increased surface ocean heat transport to the poles. This hypothesis is supported by evidence presented earlier suggesting stronger NADW formation prior to the intensification of northern hemisphere glaciation (Raymo et al 1992). Cronin (1991) also found evidence for a stronger Gulf Stream/North Atlantic Drift system at this time. Likewise, Dowsett et al (1992) proposed that the relative constancy of low-latitude SSTs in the mid-Pliocene implied stronger ocean heat transports (although the error of their method is potentially greater than the increase in SST that they were trying to exclude).

Of course, with both decreased CO_2 or decreased ocean heat transports, one must ask why these changes took place? Why would chemical weathering rates increase or why would thermohaline circulation weaken over the past few million years? These processes themselves are sensitive to climate. Chemical weathering rates can change as a function of precipitation, vegetation, or temperature. Likewise, ocean circulation patterns are controlled by evaporation-precipitation pattern as well as regional wind fields. One possibility is that increased uplift and erosional activity in mountain ranges such as the Himalaya could have resulted in a significant increase in chemical weathering rates over the Neogene (Raymo et al 1988). Sorkhabi & Stump (1993) summarize tectonic evidence for a Pliocene-Pleistocene phase of uplift and denudation in the Himalayas, and an inflection in the ocean strontium isotope record at $\sim$2.5 Ma (Capo & DePaolo 1990) also suggests an increase in chemical weathering occurred at this time. However, the possibility that these changes are due to the erosional action of mountain glaciers, which expanded in response to global cooling, can not be ruled out (Molnar & England 1990).

THRESHOLDS AND FEEDBACKS

Ultimately, the intensification of northern hemisphere glaciation which took place between 2.9 and 2.4 Ma was either a linear response to a specific change in boundary conditions taking place within this interval—such as the closing of the Panamanian Isthmus or an episode of pronounced exhumation and weathering which drew down atmospheric CO_2, or the global cooling reflected a threshold response to a longer, more gradual forcing with positive feedbacks playing an important role. For instance, a CO_2 decline occurring over a million plus years reaches a point at which perennial Arctic sea ice forms; this causes NADW formation to decrease which weakens ocean heat transport to high latitudes as well as causes a further down draw of CO_2 [e.g. via mechanisms described by Boyle (1988), Keir (1988), and/or Broecker & Peng (1989)]. Continental glaciation ultimately results.

With available data, these two scenarios are extremely difficult to distinguish. In addition, given the number of important climate feedbacks, the behavior of the system could reflect some combination of linear and nonlinear responses. Two important climate change contenders—decreases in atmospheric CO_2 and decreases in ocean heat transport to the poles—need further investigation. Application of proxy ρCO_2 methods to the Pliocene needs to be refined, as do studies of Pliocene thermohaline circulation. Lastly, the history of Arctic sea ice and its influence on deepwater formation, regional surface temperatures, and atmospheric pressure patterns requires investigation. The recent Ocean Drilling Program leg in the Arctic and Nordic Seas should provide a wealth of information on these and other topics relevant to the cause and mechanisms of northern hemisphere glaciation in the Pliocene.

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