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Cenozoic Deep-Sea Temperatures and Polar Glaciation: the Oxygen Isotope Record

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INTRODUCTION

The deep-sea paleoenvironment has undergone profound changes during the Cenozoic, including a long-term cooling of ~10°C and associated changes in ocean circulation, chemistry, and biota. During the early Eocene interval of global warmth, deep-sea temperatures reached 11-13°C, atmospheric CO_2 levels were perhaps double the present-day value, and the Earth may have been ice-free. The early Eocene is generally considered a “greenhouse world.” The transition to the “icehouse world” of the late Cenozoic involved deep-sea cooling, continental ice sheet growth on Antarctica and on the Northern Hemisphere continents, an increased planetary thermal gradient, and enhanced deep water formation in subpolar regions. Investigations into this long-term climate transition have revealed several important intervals of rapid global change, thresholds during which the climate system has shifted into a new quasi-stable state.

Identifying the nature and causes of Cenozoic climate change is critical to understanding the present climate system and predicting future global change. Emerging from continuing studies in Cenozoic paleoceanography is the hypothesis that the ocean-climate system has multiple quasi-stable states. Examining past global change will help us predict how the present climate system will evolve as we continue to perturb it. This review is intended to summarize the Cenozoic development of the present climate system. It begins with a discussion of oxygen isotopes as used in Cenozoic paleoceanography, provides an overview of deep-sea temperatures and polar ice development during the past ~70 million years (m.y.), focuses on critical intervals of extreme climates and rapid global change, and ends with a summary of late Quaternary climate variations.

Deep-sea sediments

Deciphering the Cenozoic history of the ocean-climate system relies on sediment records of past oceans obtained by deep-sea coring. Short piston coring devices typically recover 5-10 m cores, although giant piston cores may reach 40-50 m in length. Piston cores have been particularly useful in examining Quaternary ocean history. Longer sediment records from throughout the Cenozoic are accessible using drilling technology employed by the Ocean Drilling Program (ODP). With a coordinated effort by many international teams of marine geologists, ODP has made major advances in understanding the Cenozoic history of deep-sea sedimentation, ocean circulation and chemistry, and biotic evolution.

Deep-sea sediment cores are also an important source of information on the long-term history of terrestrial climates, including polar ice sheets. Because the ice sheets are in a “dynamic equilibrium," in which accumulation is generally balanced by ablation at the ice sheet margins, the present-day ice itself is quite young. For example, the age at the base of the East Antarctic ice sheet (EAIS) is probably less than ~800 thousand years (k.y.) old. Therefore, while ice core records provide detailed information on the 10^2-10^5 year scale (see Orombelli, this volume), they provide no information on ancient ice sheets on the 10^6 year scale and beyond. In contrast, information gained from deep-sea sediments can provide long-term records of Antarctic and Northern Hemisphere climate evolution over at least the course of the Cenozoic.

Benthic foraminifera from deep-sea cores have been instrumental in illuminating Cenozoic deep-sea temperatures and polar glaciation. Because benthic foraminifera inhabit the sediment-water interface, their calcite tests record bottom water conditions at the time and water depth in which they lived. By comparison, planktic foraminifera record conditions in the mixed layer and thermocline. Foraminifera from deep-sea cores therefore provide detailed records of past near-surface and bottom water conditions. In particular, oxygen isotope ratios measured on foraminiferal tests can be used to trace the evolution of ocean temperatures and global ice volume.

Oxygen Isotopes

The oxygen isotope composition of surface seawater varies predominantly as a result of evaporation/precipitation balance, in part because evaporation preferentially removes the lighter ^{16}O from seawater. Atmospheric circulation acts as a Rayleigh distillation mechanism, enriching high-latitude water vapor in ^{16}O. For these reasons, the snowfall at the South Pole reaches -50‰ (standard mean ocean water scale, SMOW), close to the average oxygen isotope composition of the Antarctic ice sheet (Shackleton & Kennett, 1975). In comparison, the $\delta^{18}O$ composition of seawater is 0‰ on the SMOW scale.

On time frames greater than the mixing time of the oceans (~1000 years; *e.g.*, Broecker & Peng, 1982), polar ice sheets therefore control the oxygen isotopic composition of seawater by preferentially storing the light isotope of oxygen (^{16}O), leaving the residual ocean enriched in ^{18}O (and hence $\delta^{18}O$). The shorthand $\delta^{18}O$ notation is as follows:

$$\delta^{18}O\ (‰) = (^{18}O/^{16}O_{sample} - {}^{18}O/^{16}O_{standard})/(^{18}O/^{16}O_{standard}) * 1000$$

Seawater is enriched in $\delta^{18}O$ during intervals when ice sheets preferentially store ^{16}O.

Oxygen isotope ratios in foraminiferal calcite are a function of the oxygen isotopic composition of seawater and the temperature of calcification. Cooler waters enrich foraminiferal calcite $\delta^{18}O$ by ~0.21‰ per °C. Calcite $\delta^{18}O$ is reported on the Pee Dee belemnite (PDB) standard scale. Estimating paleotemperatures using the temperature-dependent fractionation of calcium carbonate (Urey, 1947) has been called "one of the most striking and profound achievements in modern nuclear geochemistry" (Craig, 1965). Foraminiferal $\delta^{18}O$ values can be used to calculate ocean temperatures using a paleotemperature equation if the oxygen isotope composition of seawater is known, as in the following relation (Kim & O'Neil, 1997):

$$T°C = 16.1 - 4.64\ (\delta_C-\delta_W) + 0.09\ (\delta_C-\delta_W)^2$$

where δ_c is the $\delta^{18}O$ of calcite and δ_w is the $\delta^{18}O$ of CO_2 in equilibrium with seawater (both on the PDB scale). Present-day δ_w = -0.27‰ (Hut, 1987). Except at the lowest ocean temperatures, a linear equation can also be used (Bemis et al., 1998):

$$T°C = 16.5 - 4.8\ (\delta_C-\delta_W)$$

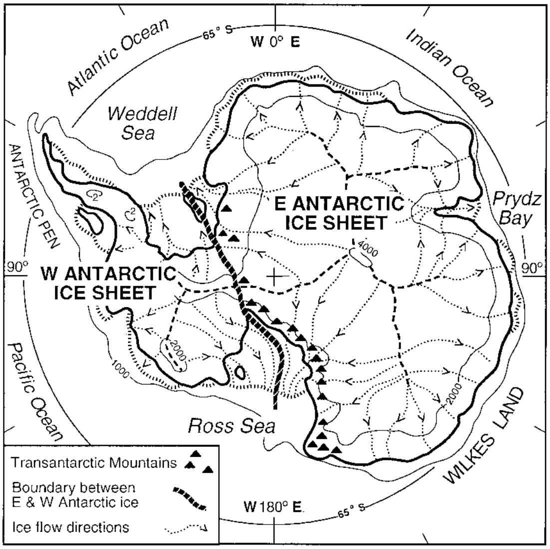


Fig. 1 - Map of the present-day Antarctic ice sheet showing division into western and eastern parts by the Transantarctic Mountains (after Barrett, 1996). Also shown are ice drainage patterns (after Drewry, 1983).

In an ice-free world, δ_w would be much lower, because ice sheets would return ^{16}O-enriched meltwater to the global ocean. The $\delta^{18}O$ of seawater (δ_w) in an ice-free world can be calculated based on the total volume of present-day continental ice and its oxygen isotope composition. With a total volume of ~30 million km^3, the Antarctic ice sheet can be divided into two unequal parts, the larger East Antarctic ice sheet (EAIS) with ~60 m of sea-level equivalent (SLE) and the smaller West Antarctic ice sheet with ~6 m SLE (Fig. 1). The Greenland ice sheet represents an additional ~6 m SLE for a total global ice volume of ~72 m SLE. Using an average $\delta^{18}O$ of -50‰ for the EAIS (Shackleton & Kennett, 1975) and an ice volume of 72 m SLE, the effect on δ_w of melting all the present-day continental ice sheets can be estimated as follows:

72 m (SLE) / 3800 m (average depth of oceans) * -50‰ = -0.95‰

Therefore in an ice-free world, δ_w would have been -0.95‰ lower than present, or -1.2‰ (PDB), and past deep-sea temperatures can be calculated readily (*e.g.*, Miller et al., 1987; Zachos et al, 1994).

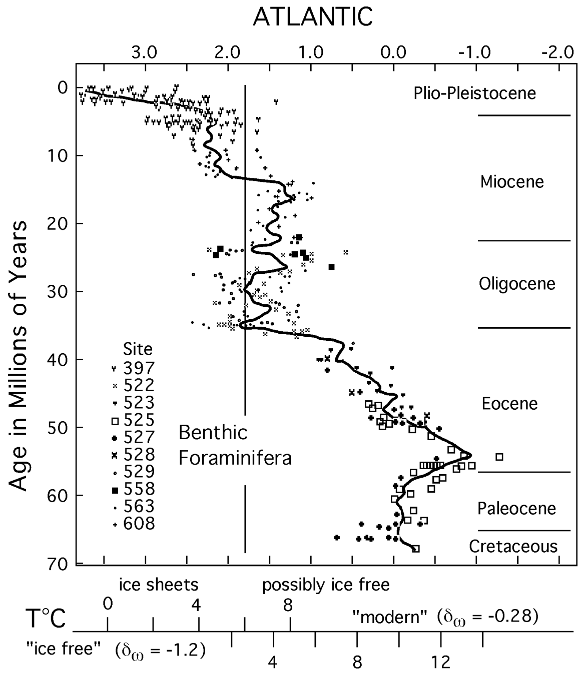
Estimating δ_w during the early Cenozoic is difficult mainly because global ice volume is unknown. Other factors such as incorporation of primordial waters from hydrothermal systems, as well as other sources, further complicate the estimate. With this *caveat*, past ocean temperatures can be calculated when δ_w is known, using the paleotemperature equation above.

CENOZOIC HISTORY

Tertiary Overview

The long-term increase in benthic foraminiferal $\delta^{18}O$ over the Cenozoic (Fig. 2) reflects some combination of deep water cooling and polar ice sheet growth. Cenozoic deep-sea temperatures were warmest during the early Eocene, reaching ~11-13°C at *ca.* 53 million years ago (Ma) (Miller et al., 1987). After the early Eocene "greenhouse world," with high sea levels and minimal continental ice sheets (including Antarctica; *e.g.*, Crowley and North, 1991; Francis, this volume), began a long-term transition toward the late Cenozoic "icehouse world." During the past ~50 m.y., deep water temperatures cooled to ~2°C and semi-permanent continental ice sheets grew, accounting for ~72 m equivalent eustatic (global sea-level) lowering. Of the long-term increase

Fig. 2 - Benthic foraminiferal oxygen isotope synthesis ($\delta^{18}O$) from Atlantic Deep Sea Drilling Project (DSDP) cores for the past 70 m.y. (from Miller et al., *Paleoceanography*, 2, 1-19, 1987, copyright by the American Geophysical Union). *Cibicidoides* $\delta^{18}O$ data are adjusted by 0.64‰ (Graham et al., 1981) for disequilibrium calcification and reported on the Pee Dee belemnite (PDB) scale. Temperature scales for modern seawater (δ_w = -0.28‰) and for an ice free world (δ_w = -1.2‰) are given. The $\delta^{18}O$ curve (smoothed with a 27-point filter) is plotted in reverse by convention, to conform with temperature scale.



in benthic foraminiferal $\delta^{18}O$ of >3‰ over the past 53 m.y., ~2.3‰ is attributable to deep-sea cooling of ~10°C while at least 0.9‰ reflects the growth of semi-permanent polar ice sheets.

Benthic foraminiferal oxygen isotope records reveal three rapid increases in $\delta^{18}O$, superimposed on a long-term increase from the early middle Eocene to present (Fig. 2). Rapid increases near the Eocene/Oligocene boundary (*ca.* 33.6 Ma), the middle Miocene (*ca.* 13.8 Ma), and the late Pliocene (*ca.* 2.7 Ma) are thought to mark intervals of polar ice sheet accumulation, as well as continued deep-sea cooling. Increased deposition of ice-rafted debris (IRD) in the Southern Ocean coincides with the two earlier inferred ice growth events, confirming expansion of the East Antarctic ice sheet (Breza & Wise, 1992; Zachos et al., 1992; Warnke et al., 1992). Glacial marine sedimentation closer to the Antarctic continent also confirms major ice sheets on East Antarctica by the earliest Oligocene (*e.g.*, Barrett, 1989; 1996; this volume). A summary of IRD occurrences documented by recent ODP legs (Fig. 3; Domack & Domack, 1991) shows that East Antarctic ice sheets preceded the initiation of Northern Hemisphere ice sheets by over 20 m.y.

Influence of plate tectonics

Major growth of the polar ice sheets was probably linked to plate tectonic changes over the course of the Cenozoic, including tectonic effects on surface- to deep ocean circulation. For example, the separation of Antarctica from Australia and South America led to the development of the circum-polar current (CPC), and the thermal isolation of Antarctica (Kennett, 1977; Kennett & Barker, 1990). Separation occurred in two main steps, the opening of the Antarctica - Australia seaway during the late Eocene, and the opening of Drake Passage during the late Oligocene. However, EAIS development was clearly underway prior to the final opening of Drake Passage, which some workers place as late as the Oligocene/ Miocene boundary (Barker and Burrell, 1977). Nevertheless, the thermal isolation of Antarctica may have been a major control on EAIS development, as well as the circum-polar current (Kennett, 1977; Kennett & Barker, 1990). Other

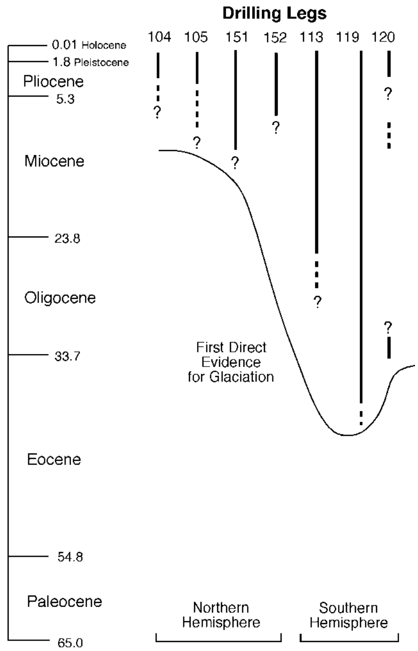


Fig. 3 - Summary of glacial ice-rafted sedimentation based on Ocean Drilling Program (ODP) legs (after Domack and Domack, 1991). Uncertain occurrences of ice-rafted debris are dashed and noted by question marks. ODP Legs 104 and 105 were located in the Norwegian Sea and Baffin Bay, respectively. Leg 151 and 152 drillsites were in Fram Strait and off the southeast Greenland margin. Legs 113, 119 and 120 were located in the Weddell Sea, Prydz Bay, and on Kerguelen Plateau, respectively, all off of East Antarctica. Cenozoic time scale is after Cande & Kent (1992) and Berggren et al. (1995).

potential controls include constriction of the eastern portal of the Tethys Ocean in the early to middle Miocene (Yilmaz, 1993). Closure of this gateway may have either decreased ocean heat transport (Woodruff & Savin, 1989), or in combination with increased North Atlantic deep water formation, increased moisture supply (Prentice & Matthews, 1991) to the high latitudes.

Tectonic controls on polar ice sheet growth may have also included sea-floor spreading rate changes and continental uplift. In addition to effects on sea level and albedo, sea-floor spreading may also affect atmospheric greenhouse gas variations. For example, intervals of fast spreading (and/or subduction) may have increased atmospheric CO_2 production and greenhouse warming (Berner, 1991). Additionally, uplift of the Tibetan Plateau and its consequent influence on orographic wind systems and chemical weathering may have enhanced Cenozoic global cooling (Ruddiman & Kutzbach, 1989; Raymo et al., 1988). Finally, uplift of the Transantarctic Mountains may have been a necessary prerequisite for major EAIS growth (see Fitzgerald, this volume; Oglesby, this volume). Plate tectonic changes, therefore, probably played a major role in the long-term evolution of polar glaciation and deep-sea temperatures.

Relation of Tertiary $\delta^{18}O$ and sea level

Miller and others (1996; 1998; Fig. 4) show good temporal agreement between Oligocene and Miocene glaciations inferred from $\delta^{18}O$ data, eustatic lowerings inferred from stratigraphic sequence boundaries, and published sea level records (Haq et al., 1987). In figure 4, a series of $\delta^{18}O$ increases is shown, from Oligocene isotope event 1 (Oi1) to Miocene isotope event 6 (Mi6), in the nomenclature of Miller et al. (1991). The earliest and largest (Oi1) is the rapid $\delta^{18}O$ increase near the Eocene/Oligocene boundary at *ca.* 33.6 Ma mentioned above. The middle Miocene $\delta^{18}O$ increase comprises the Mi3 to -4 events from *ca.* 14.0 to 12.8 Ma. The coincidence of $\delta^{18}O$ increases with New Jersey shelf-slope sequences drilled on ODP Leg 150 (Mountain et al., 1996), and with onshore New Jersey sequences (Miller et al., 1997; Fig. 4), is used to demonstrate

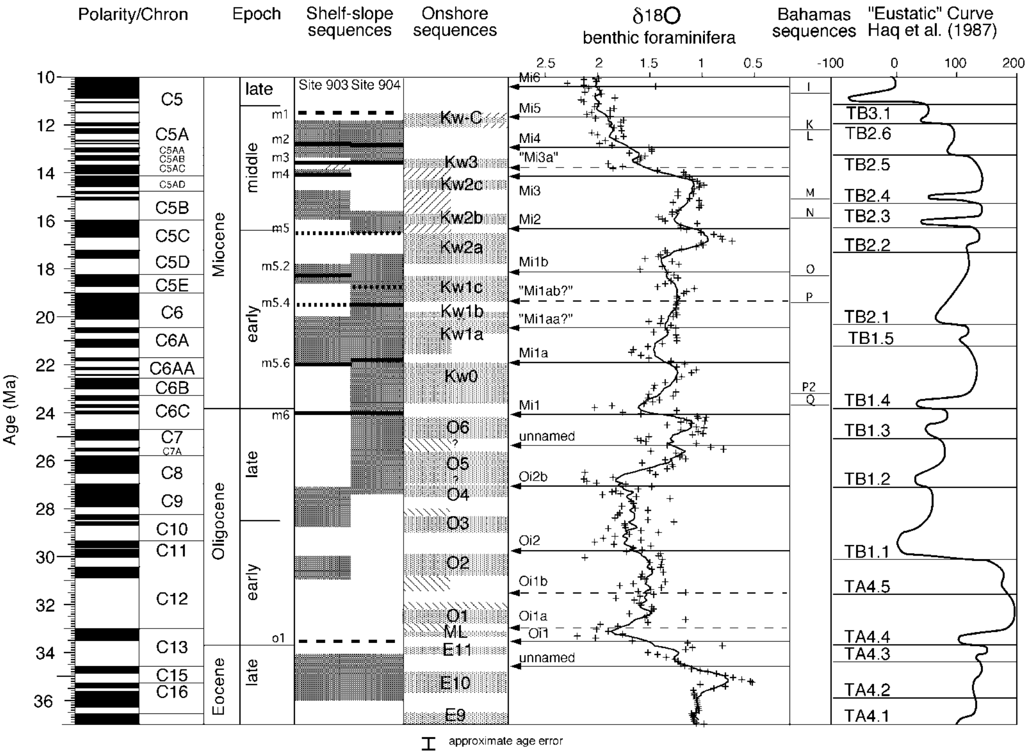


Fig. 4 - Synthesis of sea level indicators for the early Oligocene to late Miocene (from Miller et al, *Reviews of Geophysics*, 36, 569-601, 1998, copyright by the American Geophysical Union). From left to right, shelf-slope sequence boundaries from New Jersey margin, onshore sequences of New Jersey, smoothed Atlantic benthic foraminifera $\delta^{18}O$ curve with major $\delta^{18}O$ increases (after Miller et al., 1991; Wright & Miller, 1993), sequence boundaries from Barbados (Eberli et al., 1997), and "eustatic" curve of Haq et al. (1987) are compared. Major Oligocene isotope (Oi) and Miocene isotope (Mi) events are labeled (after Miller et al., 1991). Note correspondence between sequence boundaries, $\delta^{18}O$ increases, and inferred eustatic changes.

glacioeustatic changes during the Oligocene and Miocene (Miller et al., 1996; 1998). Although it is difficult to estimate the magnitude of eustatic lowering, sea level records and $\delta^{18}O$ data clearly support major glacioeustatic changes of about 50 m.

During the Eocene, benthic $\delta^{18}O$ increases and sequence boundaries show no clear correlation until the late middle Eocene (Browning et al., 1996). These observations are consistent with an ice-free "greenhouse world" during the early to mid- Eocene until ca. 42 Ma (Browning et al., 1996; Miller et al., 1998). However, significant eustatic changes throughout the Eocene (*e.g.*, Haq et al, 1987) still require explanation.

Another view holds that Antarctic ice sheets were common throughout most of the Tertiary, and may help explain sea level changes back into the Cretaceous (Matthews and Poore, 1980; Prentice and Matthews, 1988; 1991). Making the assumption that low- to mid-latitude sea-surface temperatures have remained stable throughout the Tertiary, these authors use planktic $\delta^{18}O$ from this region as a global ice volume proxy. As the western low-latitude (non-upwelling) regions (especially the western Pacific) seem to be relatively thermally stable, $\delta^{18}O$ variations can be ascribed largely to $\delta^{18}O$ compositional changes in seawater (δ_w). This interpretation of the planktic $\delta^{18}O$ record requires minimal ice growth during the middle Miocene and late Pliocene, contrary to the ice-rafted debris records. Further, the assumption of stable low- to mid-latitude sea surface temperatures may need to be re-evaluated. Some areas experienced significant glacial cooling, according to coral-based estimates in the western tropical Atlantic at the last glacial maximum (*e.g.*, Guilderson et al., 1994), although open ocean cooling may have been less.

Estimating the proportion of Cenozoic benthic $\delta^{18}O$ increase due to ice volume remains a continuing challenge. Perhaps the best approach is to evaluate the covariance between synchronous deep-sea benthic and western low-latitude planktonic $\delta^{18}O$ shifts (Shackleton & Opdyke, 1973). In this way changes in δ_w can be estimated, assuming temperature changes are minimal. Evaluating covariance during $\delta^{18}O$ increases of the Oligocene and Miocene has been used in order to estimate δ_w changes and hence ice volume changes (*e.g.*, Miller et al., 1991; Wright & Miller, 1993; Flower & Kennett, 1993). Evidence of coeval benthic and planktonic foraminiferal $\delta^{18}O$ increases confirms that δ_w changes of 0.5‰ were common at least through the Oligocene to Quaternary. Interpreting these δ_w changes requires independent information on the relation of δ_w to sea level.

A calibration of δ_w to sea level based on late Quaternary coral reef data (Fairbanks & Matthews, 1978) indicates 0.11‰ per 10 m sea level change. This value is based on late Quaternary Northern Hemisphere continental ice with an estimated $\delta^{18}O$ of -40‰. As pointed out by Miller et al. (1987), smaller Antarctic continental ice sheets in the early stages of Tertiary glaciation may have had a higher $\delta^{18}O$ composition, with values near -17‰. Therefore a calibration of 0.055‰ per 10 m sea level change may represent a useful limit for the Tertiary. A closer view of Oligocene through Miocene benthic $\delta^{18}O$ records reveals a series of large (over 0.5‰) increases (Fig. 4). During the Oligocene, covariance between benthic and planktic $\delta^{18}O$ increases is 0.3-0.9‰, reflecting eustatic lowerings of at least ~25-80 m (Miller et al., 1991; Wright & Miller, 1993). Covariance during Miocene $\delta^{18}O$ increases is 0.5-0.8‰, reflecting eustatic lowerings of a similar magnitude (Wright et al., 1992).

While the Tertiary long-term trend is one of continued ice growth and deep-sea cooling, ice growth apparently occurred in several discrete steps, which may represent new thresholds reached in the climate system. The rapidity of these steps is of great importance to understanding the interaction between continental ice sheets and the ocean-atmosphere system. Some ice growth/ deep-sea cooling events may have occurred in <1 m.y., although only a few high-resolution records exist to document the rapidity. In addition to deep-sea cooling events, there are several intervals in which deep-sea temperatures were much warmer than present, reflecting a very different climate system. In the next section, several key events of the Tertiary are examined in detail, beginning with a major warming of the deep-sea near the Paleocene/Eocene boundary.

KEY EVENTS OF THE TERTIARY

The Late Paleocene Thermal Maximum

Superimposed on the long-term trend of Cenozoic deep-sea cooling, at least one distinct warming event had profound effects on deep-sea biota. The largest extinction of benthic organisms in the past 90 m.y. (Thomas, 1990) occurred in association with a brief, but severe, warming of the deep-sea near the Paleocene/Eocene boundary (the Late Paleocene Thermal Maximum; LPTM). Oxygen isotope data based on benthic and planktic foraminifera from ODP Site 690 on Maud Rise in the Weddell Sea exhibit $\delta^{18}O$ decreases of >1.0‰ at ca. 55.33 Ma (Kennett & Stott, 1991; Fig. 5). If attributed entirely to temperature, the 2.0‰ decrease in the benthic foraminifera *Nuttalides truempyi* indicates deep-sea warming of ~8.7°C in <6 k.y to temperatures reaching ~18°C (~2100 m paleodepth). Warming based on oxygen isotope records coincides exactly with a marked decrease in benthic species richness in the same

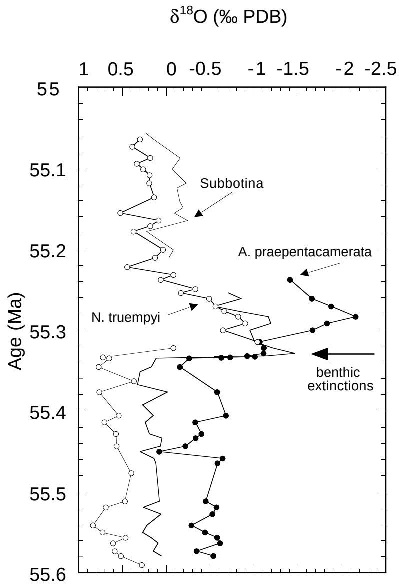


Fig. 5 - Detailed benthic and planktic oxygen isotope records across the Paleocene - Eocene boundary in ODP Site 690 on Maud Rise, Weddell Sea (after Kennett and Stott, 1991). *Nuttalides truempyi* $\delta^{18}O$ adjusted for disequilibrium calcification. Top of *Acarinina praepentacamerata* planktic $\delta^{18}O$ record represents disappearance of this taxon. The gap in the *Subbotina* $\delta^{18}O$ record occurs between the last occurrence of *S. patagonica* and the first occurrence of *S. varianta*. Note the rapid $\delta^{18}O$ decreases at ca. 55.33 Ma, coinciding with benthic extinctions, just preceding the Late Paleocene Thermal Maximum.

sediment section. The large decrease in benthic $\delta^{18}O$ led to similar benthic and planktic $\delta^{18}O$ values for a ~30 k.y. interval, indicating surface and deepwater temperatures may have been similar.

The cause of the LPTM is unknown, but may relate to fundamental changes in deep ocean circulation (Kennett and Stott, 1990; 1991), increased volcanism in subduction and/or mid-ocean ridge settings (*e.g.*, Rea et al., 1990; Bralower et al., 1997), and climatic feedbacks involving greenhouse gas release (Dickens et al., 1995; 1997). A major carbon isotope decrease coincides with the deep-sea warming, suggesting large-scale changes in carbon cycling in association with the LPTM (Kennett & Stott, 1991). The LPTM is of great interest because it may represent an extreme "greenhouse world," a transient state followed by a new quasi-stable climate.

The Eocene/Oligocene climate transition

Perhaps the largest step in the transition from the "greenhouse world" to the "icehouse world" occurred near the Eocene/Oligocene boundary (Fig. 6). This transition at ca. 33.6 Ma was clearly associated with major cooling as recorded by both marine and terrestrial organisms. A prominent benthic $\delta^{18}O$ increase of 1.3-1.4‰ (Oi1 of Miller et al., 1991) is accompanied by a well-documented appearance of IRD (Breza and Wise, 1992; Zachos et al., 1992) in the Indian sector of the Southern Ocean. Approximately 0.5‰ of the $\delta^{18}O$ increase is ascribed to ice volume increase, reflecting at least 45 m eustatic lowering. The remaining 0.9‰ is attributed to deep-sea cooling of 3-4°C, about 30-40% of total Cenozoic cooling (Zachos et al., 1994; 1996).

Assuming a bottom water temperature of 1°C during the Oi1 peak, the benthic $\delta^{18}O$ value of 2.4‰ indicates ice volumes perhaps >50% of present (Fig. 7). This estimate is a minimum, because both the assumed composition of glacial ice (-45‰) and/or the estimated temperature (1°C) could have been greater. For comparison, assuming a bottom water temperature of 4°C yields ice volumes greater than present (Zachos et al., 1994). The $\delta^{18}O$ increase occurred in <350 k.y., with the greatest change during the final 40-50 k.y. (Zachos et al., 1996), confirming a rapid climate transition during the earliest Oligocene.

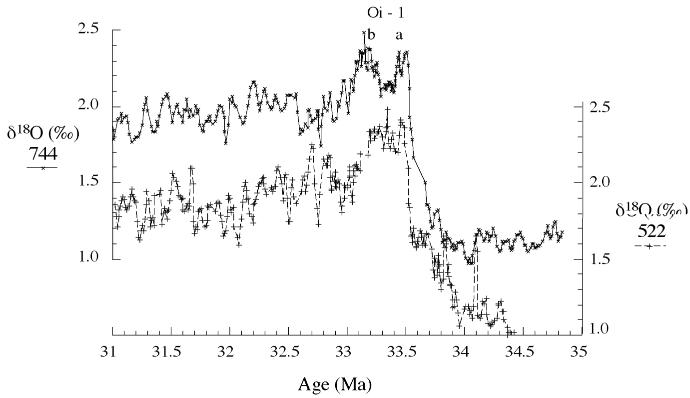


Fig. 6 - Detailed benthic foraminifera (*Cibicidoides*) $\delta^{18}O$ records from Deep Sea Drilling Project Site 532 (Angola Basin, South Atlantic) and ODP Site 744 (Kerguelen Plateau, Indian sector of Southern Ocean) across the Eocene/Oligocene boundary (from Zachos et al., *Paleoceanography*, 11, 251-266, 1996, copyright by the American Geophysical Union). Oi-1 is a large increase of 1.3-1.4‰, composed of a gradual increase followed by a rapid increase within 350 k.y. In contrast to the convention observed in other figures, $\delta^{18}O$ records are not plotted in reverse.

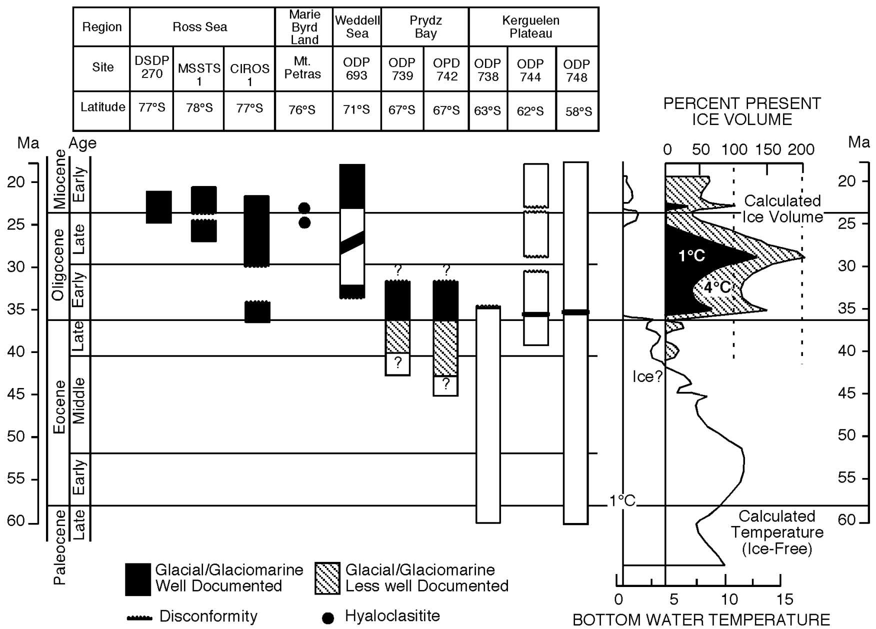


Fig. 7 - Summary of Paleogene glaciomarine sedimentation in Antarctic and southern ocean drill cores, plotted with ice volume/deep-sea temperature record computed from benthic foraminiferal $\delta^{18}O$ (from Zachos et al., *Paleoceanography*, 9, 353-387, 1994, copyright by the American Geophysical Union). Weddell Sea sites are from ODP Leg 113; Prydz Bay sites are from ODP Leg 119; Kerguelen Plateau sites are from ODP Leg 120.

Glaciations near the Oligocene/Miocene boundary

Following the Eocene/Oligocene boundary event was an interval of generally warming conditions from *ca.* 33 to 17 Ma, nevertheless punctuated by major Antarctic glaciations (Fig. 4). The high values reached during each of the $\delta^{18}O$ events, and benthic-planktic $\delta^{18}O$ covariance suggests ice volume changes of 25-80 m SLE (*e.g.*, Wright & Miller, 1993; Wright et al., 1992). The Oligocene through early Miocene glaciations cannot be understood as steps on the Cenozoic cooling trend, but rather as intermittent glaciations superimposed on a ~16 m.y. warming trend.

A recent record from the deep western equatorial Atlantic across Oligocene/Miocene boundary (Flower et al, 1997; Zachos et al., 1997) provides a detailed view of one of these intermittent glaciations, Mi1. Benthic $\delta^{18}O$ values reach 2.1‰ at ca. 23.7 Ma (Fig. 8). The Mi1 glaciation is shown to be the greatest in a series of ~400 k.y. $\delta^{18}O$ cycles, a period not often observed in Cenozoic isotope records. The $\delta^{18}O$ covaries with carbon isotope changes, suggesting carbon cycling may have enhanced climate sensitivity to Milankovitch orbital forcing in the ~400 k.y. (eccentricity) band (Flower et al, 1997; Zachos et al., 1997). Further, a strong 41 k.y. (obliquity) period within this record is consistent with high latitude forcing, including EAIS variations (see later section on Northern Hemisphere glaciation). Hence, the Mi1 glaciation is the greatest in a succession of orbitally-paced glaciations, possibly enhanced by global carbon cycling, which interrupted a long-term global warming trend. Together with high-resolution records from the Eocene/Oligocene boundary and other intervals, these findings underscore that orbitally-paced climate changes occurred during the mid-Cenozoic unipolar glacial world as well as during the Quaternary bipolar glacial world.

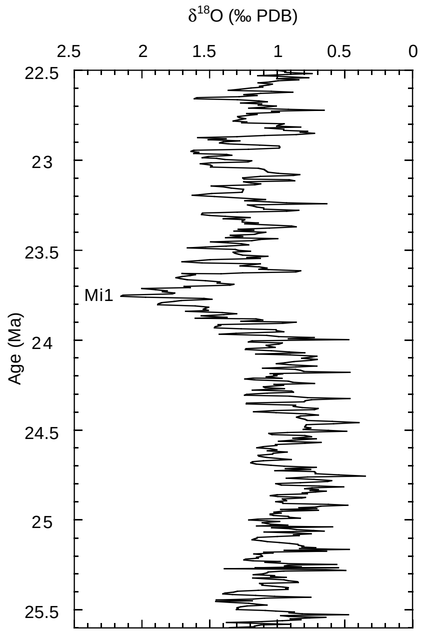


Fig. 8 - Detailed $\delta^{18}O$ record based on benthic foraminifera (*Cibicidoides*) from Ceara Rise Site 929 for the 22.5-25.5 Ma interval, including Mi1 (from Flower et al., 1997; Zachos et al., 1997). Milankovitch-scale periods are observed in the ~20-400 k.y. band.

Middle Miocene EAIS growth

The late Oligocene through early Miocene warming trend was terminated by a succession of $\delta^{18}O$ increases during the middle Miocene (Fig. 4). The largest (~1.0‰) and most rapid (<200 k.y.) was the Mi3 event at ca. 13.8 Ma (Flower and Kennett, 1993; 1995). Covariance of >0.5‰ during Mi2, -3, and -4 indicates significant eustatic lowering of at least 50 m from ca. 16 to 12 Ma (Wright et al., 1992; Flower & Kennett, 1993). An increase in IRD in Southern Ocean cores confirms expansion of the EAIS during the middle Miocene (Warnke et al., 1992). Deep-sea cooling across this interval was at least 2°C. Benthic faunal turnover (including the establishment of a foraminiferal fauna dominated by *Epistominella exigua*) in association with Mi3 is consistent with continued deep-sea cooling (Kurihara & Kennett, 1986; 1992). A synthesis on early to middle Miocene climate change is provided elsewhere (Flower & Kennett, 1994).

The middle Miocene succession of $\delta^{18}O$ increases clearly led to a new climate state, in which $\delta^{18}O$ values never returned to the low values of the Oligocene and early Miocene. Ice sheet expansion during the middle Miocene established a semi-permanent East Antarctic ice sheet. The new climatic state probably also involved establishment of the late Neogene mode of deep water circulation (Woodruff & Savin, 1989; 1991; Wright et al., 1992; Flower & Kennett, 1995). For example, benthic $\delta^{18}O$ increases were greatest at depth and in the high southern latitudes, reflecting increased production of Southern

Ocean deep water sources during this interval (Woodruff & Savin, 1991). Middle Miocene global cooling probably affected the high northern latitudes as well, perhaps preconditioning the northern hemisphere for ice sheet development.

QUATERNARY OCEAN/CLIMATE HISTORY

Northern Hemisphere glaciation

The last major benthic $\delta^{18}O$ increase of the Cenozoic occurred during the late Pliocene (Fig. 2). A detailed benthic $\delta^{18}O$ record for the mid-Pliocene to present shows a ~1‰ increase during the late Pliocene ca. 2.7 Ma (Fig. 9). The $\delta^{18}O$ increase was accompanied by widespread IRD deposition in the North Atlantic, indicative of the initiation of large-scale Northern Hemisphere glaciation (Shackleton et al., 1984). Numerous episodes of earlier ice-rafting evident in the Norwegian-Greenland Sea and Labrador Sea (*e.g.*, Fig. 3) indicate smaller, more localized ice sheets. Centers of glaciation established by the late Miocene included the Laurentide ice sheet of North America, the Greenland ice sheet, the Fennoscandian ice sheet of northwest Europe, and perhaps the Barents Sea ice sheet in the Arctic Ocean.

As discussed previously for Antarctic glaciation, plate tectonic events may have triggered Northern Hemisphere ice sheet growth. Specifically, the closure of the Panamanian Seaway during the Pliocene may have intensified the Gulf Stream and attendant transport of moisture and high salinity surface waters to the high-latitude North Atlantic, enhancing the development of the Northern Hemisphere ice sheets (*e.g.*, Maier-Reimer et al., 1990; Mikolajewicz et al., 1993; Driscoll & Haug, 1998). Alternatively, explosive volcanism may have triggered Northern Hemisphere ice growth by inhibiting solar radiation (*e.g.*, Kennett & Thunell, 1977). A major episode of subduction-related volcanism in the North Pacific at 2.75 Ma coincides with the late Pliocene benthic $\delta^{18}O$ increase, suggesting a causal relation (Rea et al., 1993). A recent review of the initiation of Northern Hemisphere glaciation and its potential causes is provided by Raymo (1994).

Superimposed upon the late Pliocene benthic $\delta^{18}O$ increase was an increase in the amplitude of glacial-interglacial cycles (Fig. 9). Spectral analysis reveals that $\delta^{18}O$ variability increased at the 41 k.y. period during the benthic $\delta^{18}O$ increase (Raymo et al., 1989). This period corresponds to variations in

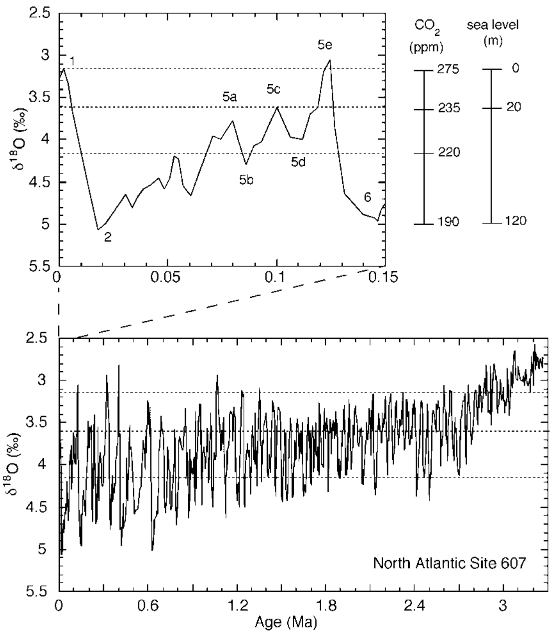


Fig. 9 - Benthic foraminifera (*Cibicidoides*) $\delta^{18}O$ record from North Atlantic Site 607 for the past 3.2 m.y. reflecting the initiation of Northern Hemisphere glaciation. Dashed horizontal lines indicate values of the Holocene, marine oxygen isotope stage (MIS) 5c, and the MIS 2/1 boundary. The last climatic cycle is expanded to show estimated sea level change and atmospheric CO_2 (from Raymo, 1994; with permission, from the *Annual Reviews of Earth and Planetary Science*, Volume 22 © by Annual Reviews).

insolation controlled by Earth's changing tilt. Because variations in Earth's tilt (from 22-24.5°) primarily affect the high latitudes, 41 k.y. variations are consistent with amplification of insolation changes by high-latitude processes. These processes include ice-sheet variations and associated changes in deepwater formation and temperatures. Hence, several lines of evidence (including IRD history in the North Atlantic and benthic $\delta^{18}O$ records) point to initiation of major Northern Hemisphere ice sheets during the late Pliocene *ca.* 2.7 Ma.

A further amplification of the $\delta^{18}O$ climate signal occurred during the mid-Quaternary (*e.g.*, Pisias & Moore, 1981; Prell, 1982; Shackleton et al., 1990; Mix et al., 1995). During the interval from *ca.* 900-700 ka, the 41 k.y. period of climatic variability was largely replaced by a 100 k.y. period (Figs. 9 and 10). The 100 k.y. period corresponds to one of the periods of eccentricity variations of the Earth's orbit, but the associated insolation changes are relatively weak. The dominance of the 100 k.y. cycle during the late Quaternary is a puzzle, but may relate to a nonlinear response of the climate system when large ice sheets are present (Imbrie et al., 1992; 1993). Therefore, the increased dominance of the 100 k.y. $\delta^{18}O$ cycle (Fig. 10) is considered to indicate a further increase in the size of the Northern Hemisphere ice sheets.

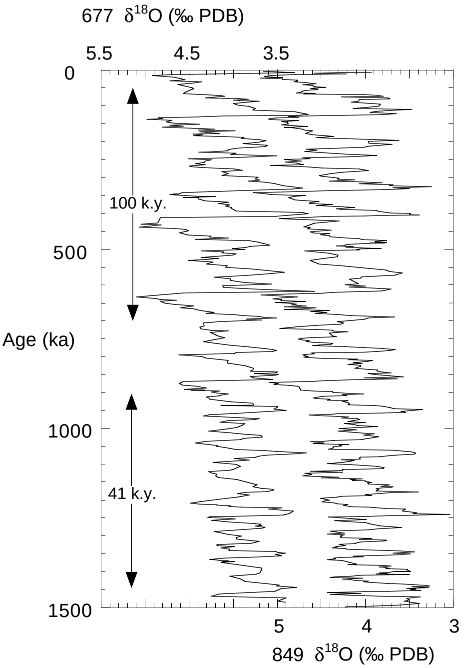


Fig. 10 - Benthic foraminifera $\delta^{18}O$ records from deep Pacific sites 677 (Shackleton and Hall, 1989; Shackleton et al., 1990) and 849 (Mix et al., 1995) for the mid- to late Quaternary. Mid-Quaternary climate transition comprising a shift in dominant period of 41 k.y. to 100 k.y. period is observed between about 900 and 700 ka.

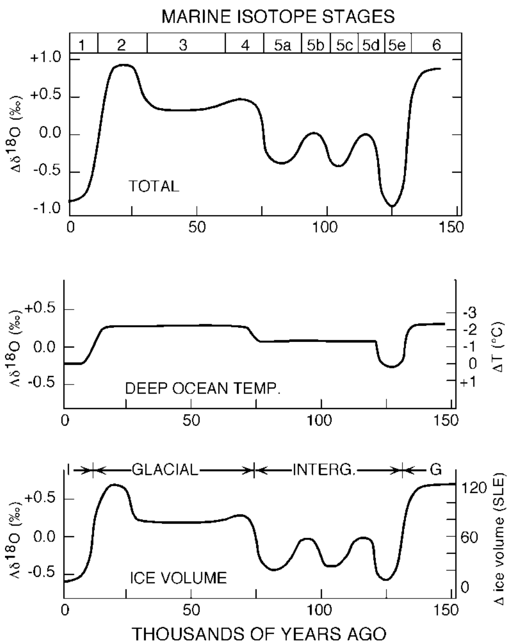


Fig. 11 - Summary of benthic foraminifera $\delta^{18}O$ record, deep ocean temperature, and global ice volume over the past 150 ka (from Broecker, 1995). All records are plotted with respect to average late Quaternary values, and equivalent temperature and ice volume, where possible. Ice volume record (panel c) derived from coral reef data is subtracted from benthic $\delta^{18}O$ record (panel a) to allow calculation of deep ocean temperatures (panel b).

Late Quaternary glacial-interglacial cycles

Late Quaternary glacial-interglacial $\delta^{18}O$ cycles were first codified by Emiliani (1955), who recognized a succession of ~2‰ oscillations in subtropical planktic $\delta^{18}O$ stretching from the present to the mid-Quaternary. Emiliani numbered these variations starting with the Holocene (oxygen isotope stage 1) such that interglaciations have odd numbers, while glaciations have even numbers (including the last glacial maximum, stage 2). This nomenclature has been formalized as “marine isotope stages” (or MIS). Figures 9-11 show some of the marine isotope stages for the mid- to late Quaternary. In his seminal 1955 work, Emiliani also correctly attributed the pacing of these glacial-interglacial $\delta^{18}O$ variations to orbital forcing, although he underestimated the proportion due to sea level change. Subsequent work confirmed that most of the benthic $\delta^{18}O$ signal was due to ice volume variations and hence indicated large sea level variations (*e.g.*, Shackleton, 1967; Shackleton & Opdyke, 1973).

Detailed information on late Quaternary sea level from coral reefs aids in determining deep-sea temperatures based on oxygen isotopes. During the last glacial maximum (LGM) at ca. 20 ka, sea level was about 120 m lower than present (Fairbanks, 1989), fixing the $\delta^{18}O$ of seawater at ~1.26‰ higher, assuming an ice sheet composition of -40‰. Sea level estimates are also available for other key intervals during the past 150 k.y., including levels similar to today during marine isotope stage (MIS) 5e (ca. 125 ka), and -20 m during MIS 5c (105 ka). Significantly, these estimates allow calibration of δ_w to ice volume and sea level (0.11‰ per 10 m sea level equivalent; Fairbanks and Matthews, 1978). Moreover, they help fix δ_w over the past 150 k.y. (Fig. 11; Broecker, 1995).

Ice volume estimates and δ_w are plotted in Figure 11c as deviations from a late Quaternary mean. Subtracting these δ_w changes from the $\delta^{18}O$ of deep Pacific benthic foraminifera (Fig. 11a) allows calculation of Pacific deep water temperatures (Fig. 11b). Temperatures were ~2°C cooler than present throughout MIS 2-4 (including the LGM), 1°C cooler than present during most of MIS 5, and similar to present during MIS 5e. These rather small deep-sea temperature changes underscore the prime importance of ice volume changes in late Quaternary $\delta^{18}O$ records. Indeed, the synchroneity of global δ_w changes due to ice volume underlies the common use of benthic $\delta^{18}O$ records for global stratigraphic correlation.

Quaternary glacial-interglacial cycles may represent the rapid switching between at least two quasi-stable modes of the climate system, an interglacial mode (such as the Holocene) and a glacial mode (such as the last glacial maximum). Glacial buildup occurs slowly relative to deglaciation, which typically takes <10 k.y., imparting a saw-tooth form to the climate signal (Figs. 9-11). Higher-frequency climate variations first documented in ice-cores and later in deep-sea cores also show this saw-tooth form, suggesting strong climatic feedbacks accelerate the deglaciation process on different time frames (also including the Late Paleocene Thermal Maximum). Global warming mechanisms are still not well understood, but records of past climate change suggest that present global change may trigger a similar acceleration of planetary temperature rise.

CONTINUING CHALLENGES

Estimation of ocean temperatures from $\delta^{18}O$ records in foraminifera requires constraints on δ_w, controlled mainly by continental ice volume variations. Deep-sea temperatures are readily calculated during global ice-free conditions (such as probably occurred during the late Cretaceous to middle Eocene) and during intervals when coral reefs fix sea level (such as the late Quaternary). In contrast, deep-sea temperature/ice volume variations are more poorly known for most of the transition from the “greenhouse world” to the “icehouse world.” Better integration of isotopic, sea level, and continental margin erosion/sedimentation records is needed to address this fundamental gap in our understanding of the climate system.

Oxygen isotope measurements in foraminifera, as well as in other carbonate-secreting organisms such as corals, offer important information in examining past climates for clues to present and future global climatic change. Oxygen isotope records have been critical in identifying and exploring intervals of extreme climatic conditions, including (among many others) the Late Paleocene Thermal Maximum (LPTM), the Eocene/Oligocene climate transition, and the last glacial maximum (LGM). A major unresolved issue is understanding how the climate system operated during these intervals. Similarly, rapid climate transitions have much to teach us about the climate system. Evidence is mounting from throughout the Cenozoic that the climate system exhibits multiple quasi-stable states, and that rapid climate transitions represent unstable, transient states between different modes of the climate system. Past climate records, therefore, offer important insights on current global change.

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REFERENCES

Barker P.F. & Burrell J., 1977. The opening of Drake Passage. *Marine Geology*, **25**, 15-34.

Barrett P.J., 1989. *Antarctic Glacial History from CIROS-1 Drill Hole, McMurdo Sound*, N.Z. DSIR Bull., **245**, 251 pp.

Barrett P.J., 1996. Antarctic Palaeoenvironment through Cenozoic Times - A review. *Terra Antartica*, **3**, 103-119.

Bemis B.E., Spero H.J., Bijma J., & Lea D.W., 1998. Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations. *Paleoceanography*, **13**, 150-160.

Berggren W.A., Kent D.V., Swisher C.C., & Aubry M.P., 1995. A revised Cenozoic geochronology and chronostratigraphy. In: Berggren W.A., Kent D.V., Aubry M.P. & Hardenbol J., (Eds.) Chronology, time scales and global stratigraphic correlation. *SEPM Spec. Pub*. **54**, 386 pp.

Berner R.A., 1991. A model for atmospheric CO_2 over the Phanerozoic time, *Am. J. Sci*., **291**, 339-376.

Bralower T.J., Thomas D.J., Zachos J.C., Hirschmann M.M., Rohl U., Sigurdsson H., Thomas E. & Whitney D.L., 1997. High-resolution records of the late Paleocene thermal maximum and circum-Caribbean volcanism: Is there a causal link? *Geology*, **25**, 963-967.

Breza J. & Wise S.W., 1992. Lower Oligocene ice-rafted debris on the Kerguelen Plateau, Evidence for East Antarctic continental glaciation. *Proc. ODP, Sci. Res*., **120**, 161-178.

Broecker W.S. & Peng T.H., 1982. *Tracers in the Sea*., Eldigio Press, New York, 690 pp.

Broecker W.S., 1995. *The Glacial World According to Wally*. Eldigio Press, New York, 330 pp.

Browning J.V., Miller K.G. & Pak D., 1996. Global implications of lower to middle Eocene sequence boundaries on the New Jersey coastal plain: The icehouse cometh. *Geology*, **24**, 639-642.

Cande S.C. & Kent, D.V., 1992. A new geomagnetic polarity time scale for the late Cretaceous and Cenozoic. *Journal of Geophysical Research*, ***97***, 13,917-19,951.

Crowley T.J. & North G.R., 1991. *Paleoclimatology*. Oxford University Press, New York, 339 pp.

Craig H., 1965. The measurement of oxygen isotope paleotemperatures. *Stable Isotopes in Oceanographic Studies and Paleotemperatures*. Spoleto, July 26-27, 1965. CNR, Laboratorio di Geologia Nucleare, Pisa, 1-24.

Dickens G.R., Castilo M.M. & Walker J.G.C., 1997. A blast of gas in the latest Paleocene: Simulating first-order effects of massive dissociation of oceanic methane hydrate. *Geology*, **25**, 259-262.

Dickens G.R., O’Neil J.R., Rea D.K. & Owen R.M., 1995. Dissociation of oceanic methane hydrate as a cause of the carbon isotope excursion at the end of the Paleocene. *Paleoceanography*, **10**, 965-971.

Domack E.W. & Domack C.R., 1991. *Cenozoic glaciation: The marine record established by ocean drilling*. JOI/USSAC, Washington, D.C., 50 pp.

Drewry D.J., 1983. *Antarctica: Glaciological and Geophysical Folio*. Scott Polar Research Institute, Cambridge, England, 9 sheets.

Driscoll, N.W. & Haug, G.H., 1998. A short circuit in thermohaline circulation: A cause for Northern Hemisphere glaciation? *Science*, **282**, 436-438.

Eberli G.P., Swart P.K., Malone M.J., et al., 1997. *Proc. ODP, Init. Reports*, **166**, College Station, TX (Ocean Drilling Program).

Emiliani C., 1955. Pleistocene paleotemperatures. *J. of Geology*, **63**, 538-578.

Fairbanks R. & Matthews R.K., 1978. The marine oxygen isotope record in Pleistocene coral, Barbados, West Indies. *Quat. Res*., **10**, 181-196.

Flower B.P. & Kennett J.P., 1993. Middle Miocene ocean/climate transition: high-resolution stable isotopic records from DSDP Site 588A, southwest Pacific. *Paleoceanography*, **8**, 811-843.

Flower B.P. & Kennett, J.P., 1994. The middle Miocene climatic transition: East Antarctic ice sheet development, deep ocean circulation, and global carbon cycling. *Palaeogeography, Palaeoclimatology, Palaeoecology,* **108**, 537-555.

Flower B.P. & Kennett J.P., 1995. Middle Miocene deep water paleoceanography in the southwest Pacific: Relations with East Antarctic ice sheet development. *Paleoceanography,* **10**, 1095-1112.

Flower B.P., Zachos J.C. & Paul H., 1997. Milankovitch-scale climate variability recorded near the Oligocene/Miocene boundary. In: Shackleton N.J., Curry W.B., Richter C. & Bralower T.J. (Eds.), *Proc. ODP, Sci. Results*, **154**, College Station, TX (Ocean Drilling Program), 433-439.

Graham D., Corliss B., Bender M. & Keigwin L.D., 1981. Carbon and oxygen isotopic disequilibria of Recent deep-sea benthic foraminifera. *Marine Micropaleontology,* **6**, 483-497.

Guilderson T.P., Fairbanks R.G. & Rubenstone J.L., 1994. Tropical temperature variations since 20,000 years ago: Modulating interhemispheric climate change. *Science*, **264**, 664-665.

Haq B.U., Hardenbol J. & Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic (250 million years ago to present). *Science,* **235**, 1156-1167.

Hut G., 1987. Stable isotope reference samples for geochemical and hydrological investigations, Report to the Director General, 42 pp., Intern. Atomic Energy Agency, Vienna.

Imbrie J., Berger A., Boyle E.A., Clemens S.C., Duffy A., Howard W.R., Kukla G., Kutzbach J., Martinson D.G., McIntyre A., Mix A.C., Molfino B., Morley J.J., Peterson L.C., Pisias N., Prell W.L., Raymo M.E., Shackleton N.J & Toggweiler, J.R., 1992. On the structure and origin of major glaciation cycles 1. Linear responses to Milankovitch forcing. *Paleoceanography*, **7**, 701-738.

Imbrie J., Berger A., Boyle E.A., Clemens S.C., Duffy A., Howard W.R., Kukla G., Kutzbach J., Martinson D.G., McIntyre A., Mix A.C., Molfino B., Morley J.J., Peterson L.C., Pisias N., Prell W.L., Raymo M.E., Shackleton N.J. & Toggweiler, J.R., 1993. On the structure and origin of major glaciation cycles 2. The 100,000-year cycle. *Paleoceanography*, **8**, 699-735.

Kennett J. P., 1977. Cenozoic evolution of Antarctic glaciation, the circum-Antarctic ocean and their impact on global paleoceanography. *Journal of Geophysical Research, **82***, 3843-3859.

Kennett J.P. & Barker P.F., 1990. Latest Cretaceous to Cenozoic climate and oceanographic developments in the Weddell Sea, Antarctic: An ocean-drilling perspective. In: Kennett J.P., Barker P.F., et al., *Proc. ODP, Sci. Results*, **113**, College Station, TX, (Ocean Drilling Program), 937-960,

Kennett J.P. & Stott L.D., 1990. Proteus and Proto-Oceanus: Ancestral Paleogene oceans as revealed from Antarctic stable isotopic results; ODP Leg 113. In: Kennett J.P., Barker P.F., et al., *Proc. ODP, Sci. Results*, **113**, College Station, TX, (Ocean Drilling Program), 865-880.

Kennett J.P. & Stott L.D., 1991. Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Palaeocene. *Nature,* **353**, 225-229.

Kennett J.P. & Thunell R.C., 1977. On explosive Cenozoic volcanism and climatic implications. *Science*, **196**, 1231-1234.

Kim S.-T. & O'Neil J.R., 1997. Equilibrium and non-equilibrium oxygen isotope effects in synthetic carbonates. *Geochimica Cosmochimica Acta*, **61**, 3461-3475.

Kurihara K. & Kennett J.P., 1986. Neogene benthic foraminifers: distribution in depth traverse. In: Kennett J.P. & von der Borch C.C., *Initial Reports of the Deep Sea Drilling Project*, **90**, 1037-1078, U.S. Government Printing Office, Washington, D.C.

Kurihara K. & Kennett J.P., 1992. Paleoceanographic significance of Neogene benthic foraminiferal changes in a southwest Pacific bathyal depth transect. *Marine Micropaleontology,* **19**, 181-199.

Maier-Reimer E., Mikolajewicz U. & Crowley T.J., 1990. Ocean GCM sensitivity experiment with an open central American isthmus. *Paleoceanography*, **5**, 349-366.

Matthews R.K. & Poore R.Z., 1980. Tertiary ^{18}O record and glacioeustatic sea-level fluctuation. *Geology,* **8**, 501-504.

Mikolajewicz U., Maier-Reimer E., Crowley T.J., & Kim K.Y., 1993. Effect of Drake and Panamanian gateways on the circulation of an ocean model. *Paleoceanography*, **8**, 409-426.

Miller K.G., Fairbanks R.G. & Mountain G.S., 1987. Tertiary oxygen isotope synthesis, sea level history, and continental margin erosion. *Paleoceanography,* **2**, 1-19.

Miller K.G., Mountain G.S., et al., 1996. Drilling and dating New Jersey Oligocene-Miocene sequences: Ice volume, global sea level, and Exxon Records. *Science,* **271**, 1092-1095.

Miller K.G., Mountain G.S., Browning J.V., Kominz M., Sugarman P.J., Christie-Blick N., Katz M.E. & Wright, J.D., 1998. Cenozoic global sea level, sequences, and the New Jersey Transect: Results from coastal plain and continental slope drilling. *Reviews of Geophysics,* **36**, 569-601.

Miller K.G., Snyder S.W., et al., 1996. *Proc. ODP, Sci. Results*, **150**, College Station, TX (Ocean Drilling Program).

Miller K.G., Wright J.D. & Fairbanks R.G., 1991. Unlocking the ice house: Oligocene-Miocene oxygen isotopes, eustacy and margin erosion. *Journal of Geophysical Research,* **96**, 6829-6848.

Mix A.C., Pisias N.G., Rugh W., Wilson J., Morey A. & Hagelberg T.K., 1995. Benthic foraminifer stable isotope record from site 849 (0-5 Ma): Local and global climate changes, *Proc. ODP, Sci. Results*, **138**, 371-412.

Mountain G.S., Miller K.G., Blum P., et al., 1996. *Proc. ODP, Sci. Results*, **150X**, College Station, TX (Ocean Drilling Program).

Pisias N.G. & Moore, T.C., 1981. The evolution of Pleistocene climate: A time series approach. *Earth and Planetary Science Letters,* **53**, 450-456.

Prell W.L., 1982. Oxygen and carbon isotope stratigraphy for the Quaternary of Hole 502B: Evidence for two modes of isotopic variability. *Init. Reports DSDP*, **68**, 455-464.

Prentice M.L. & Matthews R.K., 1988. Cenozoic ice-volume history: Development of a composite oxygen isotope record. *Geology,* **16**, 963-966.

Prentice M.L. & Matthews, R.K., 1991. Tertiary ice sheet dynamics: The snow gun hypothesis. *J. of Geophys. Res*., **96** (B4), 6811-6827.

Raymo M.E., 1994. The initiation of Northern Hemisphere glaciation. *Annual Reviews of Earth and Planetary Science*, **22**, 353-368.

Raymo M.E., Ruddiman W.F. & Froelich P.N., 1988. Influences of late Cenozoic mountain building on ocean geochemical cycles. *Geology*, **16**, 649-653.

Raymo M.E., Ruddiman W.F., Backman J., Clement B.M. & Martinson D.G., 1989. Late Pliocene variations in northern hemisphere ice sheets and North Atlantic Deep Water circulation. *Paleoceanography*, **4**, 413-446.

Rea D.K., Basov I.A., Janecek T.R. & Shipboard Scientific Party, 1993. Cenozoic paleoceanography of the North Pacific Ocean: Results of ODP Leg 145, the North Pacific transect. *EOS, Trans. Am. Geophys. Union,* **74**, 173.

Rea D., Zachos J.C., Owen R.M. & Gingerich, P.D., 1990. Global change at the Paleocene/Eocene boundary: Climatic and evolutionary consequences of tectonic events. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **79**, 117-128.

Ruddiman W.F. & Kutzbach J.E., 1989. Forcing of late Cenozoic northern hemisphere climate by plateau uplift in southern Asia and the American West, *J. Geophys. Res*., **94**, 18,409-18,427.

Shackleton N.J., 1967. Oxygen isotope analyses and Pleistocene temperatures re-assessed. *Nature*, **215**, 15-17.

Shackleton N.J., et al., 1984. Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. *Nature,* **307**, 708-711.

Shackleton N.J. & Kennett J.P., 1975. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analysis in DSDP sites 277, 279, and 281. In: Kennett J.P., Houtz R.E., et al., *Initial Reports DSDP*, **29**, Washington, D.C., (U.S. Govt. Printing Office), pp. 743-755.

Shackleton N.J. & Hall M.A., 1989. Stable isotope history of the Pleistocene at ODP Site 677. In: Becker K. Sakai H. et al., (Eds.), *Proc. ODP, Sci Results*, **111**, College Station, TX (Ocean Drilling Program), 295-316.

Shackleton N.J. & Opdyke N.D., 1973. Oxygen isotopic and palaeomagnetic stratigraphy of equatorial Pacific piston core V28-238: Oxygen isotope temperatures and ice volume on a 10^5 and 10^6 year scale. *Quat. Res*., **3**, 39-55.

Shackleton N.J., Berger A. & Peltier, W.R., 1990. An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677, *Trans. R. Soc. Edinburgh Earth Sci*., **81**, 251-261.

Thomas E., 1990. Late Cretaceous - early Eocene mass extinction in the deep-sea. In: Sharpton V.L. & Ward P.D. (eds.), *Global Catastrophes*, GSA Special Publication **247**, 481-496.

Urey H.C., 1947. The thermodynamic properties of isotopic substances, *J. Chem. Soc. (London),* 562-581.

Warnke D.A., Allen C.P., Muller D.W., Hodell D.A. & Brunner C.A., 1992. Miocene-Pliocene Antarctic glacial evolution: A synthesis of ice-rafted debris, stable isotope, and planktonic foraminiferal indicators, ODP Leg 114. In: Kennett J.P. & Warnke D.A., (eds.), *The Antarctic Paleoenvironment: A Perspective on Global Change, Part I.* Antarctic Research Series, American Geophysical Union, 56:311-325.

Woodruff F. & Savin S.,1989. Miocene deepwater oceanography. *Paleoceanography,* **4**, 87-140.

Woodruff F. & Savin S., 1991. Mid-Miocene isotope stratigraphy in the deep sea: high-resolution correlations, paleoclimatic cycles, and sediment preservation. *Paleoceanography,* **6**, 755-806.

Wright J.D., Miller K.G. & Fairbanks, R.G., 1992. Early and middle Miocene stable isotopes: implications for deepwater circulation and climate. *Paleoceanography,* **7**, 357-389.

Wright J.D. & Miller K.G., 1993. Southern Ocean influences on late Eocene to Miocene deep-water circulation. In: Kennett J.P. & Warnke D.A., (Eds.), *The Antarctic Paleoenvironment: A Perspective on Global Change, Part II.* Antarctic Research Series, American Geophysical Union, **60**, 1-25.

Zachos J.C., Breza J. & Wise S.W., 1992. Early Oligocene ice-sheet expansion on Antarctica: Sedimentological and isotopic evidence from Kerguelen Plateau. *Geology,* **20**, 569-573.

Zachos J.C., Flower B.P. & Paul H., 1997. Orbitally paced climate oscillations across the Oligocene/Miocene boundary. *Nature*, **388** 567-570.

Zachos J.C., Stott L.D. & Lohman K.C., 1994. Evolution of early Cenozoic marine temperatures. *Paleoceanography* **9**, 353-387.

Zachos J.C., Quinn T.M. & Salamy K.A., 1996. High-resolution (104 years) deep-sea foraminiferal stable isotope records of the Eocene-Oligocene climate transition. *Paleoceanography* **11**, 251-266.

Yilmaz Y., 1993. New evidence and model on the evolution of the southeast Anatolian orogen. *Geological Society of America Bulletin,* **105**, 251-271.