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

B.P. Flower

Department of Marine Science, University of South Florida, St. Petersburg, FL 33701 - U.S.A.
(bflower@seas.marine.usf.edu)

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 CO₂ 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 $10^2$-$10^3$ 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:

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

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 $\delta_c$ is the $\delta^{18}$O of calcite and $\delta_w$ is the $\delta^{18}$O of CO$_2$ in equilibrium with seawater (both on the PDB scale). Present-day $\delta_w = -0.27‰$ (Hut, 1987). Except at the lowest ocean temperatures, a linear equation can also be used (Bemis et al., 1998):

$$T^°C = 16.5 - 4.8 (\delta_c - \delta_w)$$
In an ice-free world, $\delta_w$ would be much lower, because ice sheets would return $^{16}$O-enriched meltwater to the global ocean. The $\delta^{18}$O of seawater ($\delta_w$) 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 $\delta_w$ of melting all the present-day continental ice sheets can be estimated as follows:

$$72 \text{ m (SLE)} / 3800 \text{ m (average depth of oceans)} \times -50‰ = -0.95‰$$

Therefore in an ice-free world, $\delta_w$ 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 $\delta_w$ 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 $\delta_w$ 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
in benthic foraminiferal δ¹⁸O 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 δ¹⁸O, 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
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 δ$^{18}$O and sea level

Miller and others (1996; 1998; Fig. 4) show good temporal agreement between Oligocene and Miocene glaciations inferred from δ$^{18}$O data, eustatic lowerings inferred from stratigraphic sequence boundaries, and published sea level records (Haq et al., 1987). In figure 4, a series of δ$^{18}$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 δ$^{18}$O increase near the Eocene/Oligocene boundary at ca. 33.6 Ma mentioned above. The middle Miocene δ$^{18}$O increase comprises the Mi3 to -4 events from ca. 14.0 to 12.8 Ma. The coincidence of δ$^{18}$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 δ¹⁸O curve with major δ¹⁸O 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, δ¹⁸O 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 δ¹⁸O data clearly support major glacioeustatic changes of about 50 m.

During the Eocene, benthic δ¹⁸O 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 δ¹⁸O 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, δ¹⁸O variations can be ascribed largely to δ¹⁸O compositional changes in seawater (δw). This interpretation of the planktic δ¹⁸O 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
Cenozoic Deep-Sea Temperatures and Polar Glaciation: the Oxygen Isotope Record

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 δ¹⁸O 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 δ¹⁸O shifts (Shackleton & Opdyke, 1973). In this way changes in δw can be estimated, assuming temperature changes are minimal. Evaluating covariance during δ¹⁸O increases of the Oligocene and Miocene has been used in order to estimate δw 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 δ¹⁸O increases confirms that δw changes of 0.5‰ were common at least through the Oligocene to Quaternary. Interpreting these δw changes requires independent information on the relation of δw to sea level.

A calibration of δw 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 δ¹⁸O 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 δ¹⁸O 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 δ¹⁸O records reveals a series of large (over 0.5‰) increases (Fig. 4). During the Oligocene, covariance between benthic and planktic δ¹⁸O 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 δ¹⁸O 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 δ¹⁸O 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
sediment section. The large decrease in benthic δ¹⁸O led to similar benthic and planktic δ¹⁸O 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 δ¹⁸O 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 δ¹⁸O 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 δ¹⁸O 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 δ¹⁸O 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.
Following the Eocene/Oligocene boundary event was an interval of generally warming conditions from \textit{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 δ¹⁸O 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. δ¹⁸O cycles, a period not often observed in Cenozoic isotope records. The δ¹⁸O 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.

**Middle Miocene EAIS growth**

The late Oligocene through early Miocene warming trend was terminated by a succession of δ¹⁸O 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 δ¹⁸O increases clearly led to a new climate state, in which δ¹⁸O 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 δ¹⁸O 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 $\sim 1\%$ 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 δ¹⁸O records) point to initiation of major Northern Hemisphere ice sheets during the late Pliocene ca. 2.7 Ma.

A further amplification of the δ¹⁸O 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. δ¹⁸O cycle (Fig. 10) is considered to indicate a further increase in the size of the Northern Hemisphere ice sheets.

Fig. 10 - Benthic foraminifera δ¹⁸O 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 δ¹⁸O 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 δ¹⁸O record (panel a) to allow calculation of deep ocean temperatures (panel b).
Late Quaternary glacial-interglacial cycles

Late Quaternary glacial-interglacial δ¹⁸O cycles were first codified by Emiliani (1955), who recognized a succession of ~2‰ oscillations in subtropical planktic δ¹⁸O 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 δ¹⁸O variations to orbital forcing, although he underestimated the proportion due to sea level change. Subsequent work confirmed that most of the benthic δ¹⁸O 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 δ¹⁸O 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 δw to ice volume and sea level (0.11‰ per 10 m sea level equivalent; Fairbanks and Matthews, 1978). Moreover, they help fix δw over the past 150 k.y. (Fig. 11; Broecker, 1995).

Ice volume estimates and δw are plotted in Figure 11c as deviations from a late Quaternary mean. Subtracting these δw changes from the δ¹⁸O 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 δ¹⁸O records. Indeed, the synchronicity of global δw changes due to ice volume underlies the common use of benthic δ¹⁸O 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 δ¹⁸O records in foraminifera requires constraints on δw, 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.

ACKNOWLEDGMENTS

This contribution was presented in the “Geological Records of Global and Planetary Changes” workshop at the International School of Earth and Planetary Sciences, University of Siena, April 1998. I thank P.J. Barrett and G. Orombelli for inviting my participation, P.J. Barrett and R.J. Oglesby for reviews of the chapter, C.A. Ricci and J. Müller for organizing the workshop, and C. Edmisten for graphics assistance. Permission to use published figures was graciously provided by W.S. Broecker, C.R. Domack, E.W. Domack, J.P. Kennett, K.G. Miller, M.E. Raymo, and J.C. Zachos. Finally, I thank the students of the International School for their interest and enthusiasm. This work was supported by NSF (EAR-9725311).

REFERENCES


