Middle Oligocene cooling from equatorial Pacific DSDP Site 77B

Lloyd Keigwin
Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, 02543

Gerta Keller
U.S. Geological Survey, Menlo Park, California 94025 and Department of Geology, Stanford University, Stanford, California 94305

ABSTRACT

An apparently complete 150-m Oligocene sequence of pelagic sediment at equatorial Pacific Deep Sea Drilling Project Site 77B has been analyzed for stable-isotope and quantitative planktonic foraminiferal studies. Lower abundance of inferred warm-water species and enrichment of ^{18}O and ^{12}C in foraminiferal tests during the interval 29 to 32 m.y. ago suggest cooler climate, lowered sea level, and significant volume of continental ice compared to earlier and later times in the Oligocene.

INTRODUCTION

Studies of continental margins suggest a dramatic lowering of sea level in Oligocene time (Quilty, 1977; Pitman, 1978; Vail and Hardenbol, 1979; Olsson et al., 1980; Siesser and Dingle, 1981; Loutit and Kennett, 1981), although the timing of this event is sometimes in dispute. For example, Vail and Hardenbol (1979) found that the greatest sea-level fall of the Cenozoic occurred about 29 m.y. ago, but on the coastal plain of Maryland, Delaware, and New Jersey, Olsson et al. (1980) found that a large regression occurred prior to 32 m.y. ago. Whereas transgressions and regressions may result from small changes in the rate of sea-level change (Pitman, 1978), large abrupt global changes in sea level are most easily explained by growth or decay of continental ice (Vail et al., 1977; Matthews and Poore, 1980). Hence, large changes in continental ice volume should be evident in oxygen-isotope records of deep-sea cores.

In Quaternary deep-sea stable-isotope studies, the strongest evidence for a seawater compositional effect due to ice volume is the covariance of benthic and planktonic foraminiferal $\delta^{18}O$ (Shackleton, 1967; Shackleton and Opdyke, 1973; Crowley and Matthews, 1983). Interpretations of pre-Quaternary stable-isotope results differ depending on at what point in time significant accumulation of continental ice is assumed. In the middle 1970s it was widely assumed that significant continental ice did not accumulate before ~15 m.y. ago, (Savin et al., 1975; Shackleton and Kennett, 1975). In 1980, however, Matthews and Poore challenged that view and proposed that ice sheets occurred earlier in the Cenozoic. This alternative interpretation assumed tropical sea surface temperatures similar to today, with major ice accumulation in the Paleogene and stepwise cooling of bottom waters through the Cenozoic (Matthews and Poore, 1980).

Interpreting the significance of continental ice volume in pre-Quaternary records is difficult because the amplitude of pre-Quaternary $\delta^{18}O$ signals is about one-half that known from the Quaternary. Thus, whereas in the Quaternary 2‰ changes in deep sea benthic foraminiferal $\delta^{18}O$ require ice volume, because otherwise bottom waters would freeze, smaller isotopic changes in the Paleogene can be explained entirely by temperature. Nevertheless, ice-volume effects are suggested whenever there is benthic-planktonic foraminiferal covariance in $\delta^{18}O$. For example, covariance indicates that the $\delta^{18}O$ increase of 1‰ at the Eocene-Oligocene boundary may be one-third ice-volume increase and two-thirds temperature decrease (Keigwin, 1980). Although Vail et al. (1977) reported a small fall in sea level, Loutit et al. (1983) have reported a rise in sea level at

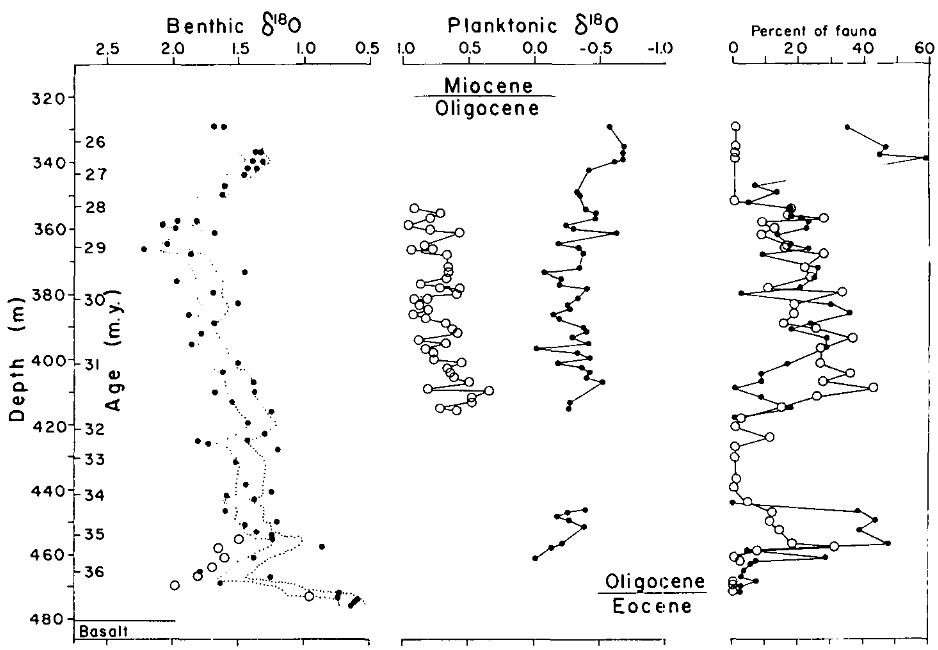

Figure 1. Oxygen-isotope results on Oligocene benthic and planktonic foraminifera and abundance of two planktonic species, DSDP Site 77B. Benthic foraminifera: open circles = *Oridorsalis tener*; solid dots = *Cibicidoides* spp. Planktonic foraminifera: open circles = *G. opima*; solid dots = *G. angustiumbilicata*. Significant glaciation ~29 m.y. ago is suggested by peak in benthic foraminiferal $\delta^{18}O$.

 GEOLOGY, v. 12, p. 16–19, January 1984

the Eocene-Oligocene boundary. Nevertheless, the major fall at 29 m.y. ago is still the dominant feature of the Cenozoic. Convincing oxygen-isotope evidence for glaciation 29 m.y. ago has not previously been published, probably because of too few data from long complete sequences. Several studies, however, have published curves with one-point $\delta^{18}O$ peaks in the middle–late Oligocene (Savin et al., 1975; Shackleton and Kennett, 1975; Boersma and Shackleton, 1977; Vergnaud Grazzini and Rabussier Lointier, 1980; Poore and Matthews, 1984). These ^{18}O-rich intervals approach in magnitude the Eocene-Oligocene ^{18}O enrichment and suggest continental glaciation.

DSDP SITE 77B

We chose Deep Sea Drilling Project (DSDP) Site 77B for a first-order study of Oligocene stable isotopic and planktonic foraminiferal variability. Site 77B lies at a water depth of 4,291 m, at 00°28.9′N and 133°13.7′W, the westernmost of an equatorial transect of sites drilled on leg 9 (Hays, Cook, et al., 1972). Severe drilling disturbance was confined to the upper 100 m of the section; the Oligocene section, from ~320 to 470 m, consists of a generally undisturbed unit of massive foraminiferal-nannofossil chalk and foraminiferal-radiolarian-nannofossil chalk. Core recovery averaged 91% for our study interval, and the average sedimentation rate was ~1 cm/1,000 yr (Hays, Cook et al., 1972). During the Cenozoic, seafloor at the location of Site 77 drifted northward and westward. Northward drift during Oligocene time was about 2°, but the site remained south of the equatorial zone of high productivity (Van Andel et al., 1975).

Preservation of foraminifera is good to excellent throughout the Oligocene, although there are some intervals, especially within the *Globigerina ampliapertura* zone, where foraminifera are few because of carbonate dissolution or dilution by siliceous microfossils. Uppermost Eocene sediments, overlying basalt basement, show evidence of high-temperature alteration (Hays, Cook et al., 1972), and planktonic foraminifera are absent.

METHODS

Samples of 20-cm^3 size were taken from Site 77, cores 36–52, every 1.5 m. Samples were wet-sieved at 73 μm and dried at 50 °C. Splits of 300–500 specimens of planktonic foraminifera of the size fraction >149 μm were counted and identified for biostratigraphic and quantitative faunal analyses. The planktonic species *Globigerina angustiumbilicata* was picked from the size fraction 149–180 μm and *Globorotalia opima* from the size fraction 212–300 μm. The benthic foraminifera genus *Cibicidoides* was picked from the >150 μm fraction for isotopic analysis. Stable-isotope procedures were similar to those described elsewhere (Keigwin, 1979), with the following modifications: samples were sonicated in methanol, benthic foraminifera were crushed in methanol prior to roasting at 370 °C, and volatiles were condensed with liquid nitrogen as they were produced in H_3PO_4.

RESULTS AND DISCUSSION

Biostratigraphy

On the basis of our faunal observations, we have applied a foraminifera zonation (Bolli, 1957; Blow, 1969) as described in Keller (1983), and applied a nannofossil CP zonation (Okada and Bukry, 1980) based on the results of Bukry (1972). These zonations are shown in Figure 2; in both Figure 1 and Figure 2 we have assigned absolute ages from the latest chronostratigraphy of Berggren et al. (1984).

Both our stable-isotope results (see below) and those of Poore and Matthews (1984) from the South Atlantic suggest that *G. angustiumbilicata* was a warm-water, surface-dwelling species, because it has low oxygen-isotope values. Faunal abundance results in Figure 1 show major peaks in abundance of this species between about 34 and 35 m.y. ago and in the interval younger than about 28 m.y. ago that we attribute to warmer surface waters. The general scarcity of this species and other inferred surface and intermediate dwellers between ~32–34 m.y. ago is due to increased $CaCO_3$ dissolution and/or dilution by silica (Keller, 1983). Surface waters may have been cooler between ~32 and 28 m.y. ago, as shown by greater abundance of the cooler water (isotopically heavier) *G. opima* (Fig. 1).

Stable Isotopes

Although we took one sample per core section (about every 1.5 m), not all samples had enough specimens for isotopic analysis. Thus, at a section with low sedimentation rate, like Site 77B, our minimum sampling interval is ~150,000 yr, which is not adequate to resolve climatic change of the frequency seen in the Quaternary. However, lower frequency change (on the order of 10^6 yr) is evident in Site 77B results (tabulated by L. Keigwin and B. H. Corliss, in prep.) Considering our wide sampling intervals, we emphasize general trends rather than individual data points; for this reason we plot as dotted lines ±0.1‰ limits about a three-point running average of benthic foraminiferal data (Figs. 1, 2).

The Oligocene begins at Site 77B with a 1‰ increase in benthic foraminiferal $\delta^{18}O$ over latest Eocene values (Fig. 1). Although latest Eocene foraminifera at Site 77B are discolored and possibly recrystallized, early Oligocene and younger foraminifera appear well

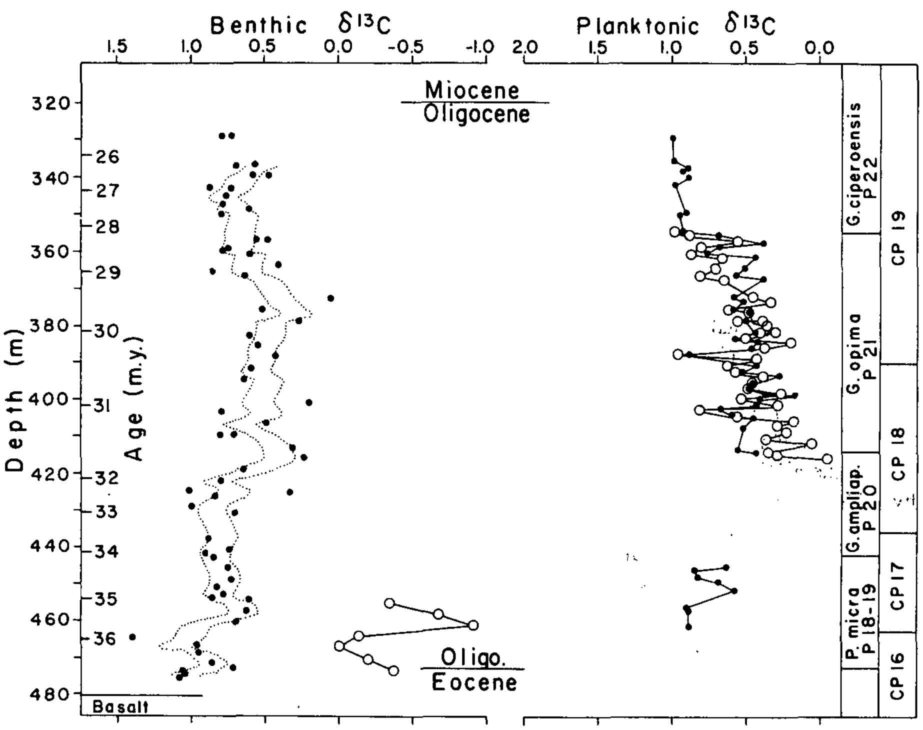

Figure 2. Carbon-isotope results and biostratigraphic zonation (see text), DSDP Site 77B. Symbols as in Figure 1. Lower $\delta^{13}C$ is seen in foraminifera of 29- to 32-m.y. age. This corresponds to interval of inferred ice growth and may be response to lowered sea level.

preserved and the earliest Oligocene $\delta^{18}O$ maximum is a feature that is seen in many detailed studies of cores of that age (Kennett and Shackleton, 1976; Keigwin, 1980; Miller and Curry, 1982; L. Keigwin and B. H. Corliss, in prep.). In the western equatorial Pacific, planktonic foraminifera show some ^{18}O enrichment across the Eocene-Oligocene boundary, but the magnitude of the change in surface dwellers at low latitudes is less than that in benthic foraminifera. This observation resulted in the previous suggestion that most of the Eocene-Oligocene change reflects lowered temperature (Keigwin, 1980).

From early Oligocene values of ~1.3‰, the $\delta^{18}O$ of *Cibicidoides* spp. increases, beginning about 32 m.y. ago (Fig. 1). During this interval of ^{18}O enrichment, $\delta^{13}C$ of both benthic and planktonic foraminifera is lowered by as much as 0.5‰ (Fig. 2). In the late Oligocene, ~28 m.y. ago, the $\delta^{18}O$ of benthic and planktonic foraminifera decreases, and their $\delta^{13}C$ increases.

Our oxygen-isotope results do not allow a unique interpretation; some combination of change in Oligocene ice volume and temperature is indicated. Two arguments favor significant continental ice volume in Oligocene time. The first is based on nearly synchronous decreases in $\delta^{18}O$ values of benthic and planktonic foraminiferal calcite in the latest Oligocene; this suggests a seawater $\delta^{18}O$ change possibly resulting from a decrease in continental ice (Fig. 1). The magnitude of change common to both benthics and plankton(~0.4‰) is comparable to that observed at the Eocene-Oligocene boundary farther to the west at DSDP Site 292 (Keigwin, 1980). If continental ice decreased in the late Oligocene, then it must have been present in the middle Oligocene, although it need not have reached a maximum at 29 m.y. ago.

Middle Oligocene benthic foraminiferal $\delta^{18}O$ values of 2‰ are stronger evidence for significant continental ice in Oligocene time (Fig. 1). Holocene *Cibicidoides* spp. from Pacific deep-sea cores typically have $\delta^{18}O$ values of ~2.5‰ (see, for example, results of Graham et al., 1981). If all continental ice today was melted, the $\delta^{18}O$ of seawater would decrease by ~0.9‰ (Shackleton and Kennett, 1975), so *Cibicidoides* spp. would register a $\delta^{18}O$ of ~1.6‰, assuming bottom-water temperatures as cold as today. We estimate that in Oligocene time Site 77B was at a depth of ~3 km, which today has a temperature of ~1.8 °C. Oligocene $\delta^{13}O$ results of ~2‰ would require temperatures in the Oligocene Pacific (at ~3 km) of ~2 °C cooler, or ~-0.2 °C assuming *Cibicidoides* fractionated ^{18}O the way it does today. Such cold temperatures at ~3 km paleodepth are unlikely, because the thermal structure of the Oligocene ocean is thought to be similar to today (Douglas and Savin, 1978). If deep-water temperatures were <0 °C, surface temperatures at least that low would be required in the source region for the water mass. Such low temperatures could easily support continental ice provided there was a source of moisture. The most reasonable interpretation of large benthic $\delta^{18}O$ values at ~29 m.y. ago is that there was a peak in continental ice volume.

Oxygen-isotope results from planktonic foraminifera are ambiguous (Fig. 1). The shallowest dwelling (isotopically lightest) species *G. angustiumbilicata* does not appear to record the inferred middle Oligocene ice-volume growth, even though it does appear to record its decay in the latest Oligocene. Deeper-dwelling *G. opima* of 31–32 m.y. ago do appear to record a slight ^{18}O enrichment, but not as much as the benthic foraminifera. Unfortunately, both these species are too rare for isotopic analysis in the critical interval 32–34 m.y. where benthic foraminifera are isotopically lightest. We speculate that small local changes in surface-water temperature and salinity may have occurred during the inferred growth of continental ice. For example, a 2 °C warming of surface waters could mask a seawater compositional effect of 0.5‰.

Oligocene $\delta^{13}C$ changes at Site 77B are useful in interpreting $\delta^{18}O$ results because they may share a common cause. Low $\delta^{13}C$ has been observed in middle Oligocene foraminifera at other locations, indicating that Site 77 results may reflect an oceanwide phenomenon. The earlier studies of Savin et al. (1975) and Shackleton and Kennett (1975) did not specifically address the Oligocene, but each, with limited sampling, found low $\delta^{13}C$ values in the Oligocene at Pacific locations. Additional data confirm this observation in the Atlantic (Boersma and Shackleton, 1977; Miller and Curry, 1982; L. Keigwin and B. H. Corliss, in prep.). Since Site 77B results clearly show decreased planktonic and benthic $\delta^{13}C$ values in the middle Oligocene, the $\delta^{13}C$ of CO_2 probably changed oceanwide.

Lowered sea level may be a mechanism for the lowering of the $\delta^{13}C$ of CO_2. Sea level is thought to have been lowered in the middle Miocene and late Miocene (Vail et al., 1977) probably because of increased continental glaciation. Woodruff et al. (1981) showed lowered benthic foraminiferal $\delta^{13}C$ results in the middle Miocene, when $\delta^{18}O$ results increase because of inferred glacial advance (Savin et al., 1975; Shackleton and Kennett, 1975). Benthic and planktonic foraminiferal $\delta^{13}C$ decreases for the late Miocene and has been specifically associated with sea-level fall due to glaciation (Loutit and Kennett, 1979; Vincent et al., 1980; Loutit and Keigwin, 1982). Lowered carbon-isotope ratios have also been noted during glacial advances of the Pleistocene (Shackleton, 1977; Broecker, 1982). In the Shackleton (1977) model, transfer of ^{12}C to the ocean occurs as a consequence of climatic cooling and reduced vegetation, especially in tropical rainforests. Broecker (1982) proposed that the observed lowering of benthic $\delta^{13}C$ during glaciations may result from either an increase in the phosphorus to carbon ratio in the deep sea through oxidation of organic matter on continental shelves exposed by lowered sea level, or by an increased ratio of phosphorus to carbon in organic matter falling from surface water to the deep sea. Although glacial advances, lowered sea level, and lowered $\delta^{13}C$ during linked, the details of this connection must await further study.

Vail et al. (1977) suggested that the middle Oligocene sea-level fall was the largest of the Cenozoic. This observation may be reconciled with our relatively small Oligocene $\delta^{18}O$ changes (compared to an estimated 1.2‰ ice volume effect in the Quaternary; Duplessy, 1978) in two ways. First, the Vail et al. (1977) curve may be influenced by local or regional conditions. For example, Loutit and Kennett (1981) suggested that the apparent magnitude of a sea-level fall could be exaggerated by the position of the shoreline on a continental margin. With the shoreline close to the shelf edge, a fall in sea level would appear larger. This explanation would imply that sea level was already significantly lowered prior to the inferred maximum glaciation at ~29 m.y. ago and is consistent with our observation of increasing benthic foraminiferal $\delta^{18}O$ values beginning ~32 m.y. ago (Fig. 1). Second, ice accumulating on continents may not have been as enriched in ^{16}O during Oligocene time as it is in the Quaternary. Because it seems most likely that ice accumulated in polar regions and because the $\delta^{18}O$ of precipitation is at least partly a function of temperature of precipitation (Dansgaard, 1964), then it follows that with a pole-to-equator gradient of air temperature lower than today, enough ice could be stored to cause large sea-level changes with smaller seawater compositional changes. Limited evidence indicates that the planetary temperature gradient was less in pre-Quaternary time (Savin et al., 1975). Thus, our 29-m.y.-ago $\delta^{18}O$ maximum could really reflect a very large sea-level fall. Unfortunately, smaller seawater compositional changes due to isotopically heavier continental ice make it more difficult to detect benthic-planktonic foraminiferal covariance in $\delta^{18}O$, which in the Quaternary is taken as the best evidence of the ice-volume–$\delta^{18}O$ effect (Shackleton and Opdyke, 1973).

SUMMARY

Although we cannot yet separate ice-volume and temperature effects in stable-isotope records, clearly Earth's climate was more severe between about 28 and 31 m.y. ago and milder before and after. Most severe conditions occurred about 29 m.y. ago. Faunal abundance changes suggest cooler surface waters during the interval 28 to 32 m.y. ago. Oxygen isotopes indicate that there was probably significant continental ice cover with a peak volume at ~29 m.y. ago, in agreement with seismic stratigraphic evidence for widespread regression. If there was no continental ice, surface-water temperature at high latitudes was probably below 0 °C, which would promote ice growth provided there was a moisture source. Continental ice was probably more enriched then in ^{18}O than today, possibly because of a lower planetary temperature gradient, thus accounting for relatively small $\delta^{18}O$ changes and large sea-level changes. Similar to other times in the Tertiary and Quaternary, carbon-isotope results are light during the interval of inferred climatic cooling and regression.

REFERENCES CITED

Berggren, W. A., Kent, D. V., and Flynn, J. J., 1984, Paleogene geochronology and chronostratigraphy, *in* Snelling, N. J., ed., Geochronology and the geological record: Geological Society of London Special Paper (in press).

Blow, W. H., 1969, Late middle Eocene to Recent planktonic foraminiferal biostratigraphy, *in* Proceedings, First International Conference of Planktonic Microfossils, Geneva, 1967, Volume 1: Leiden, E. J. Brill, p. 199–422.

Boersma, A., and Shackleton, N. J., 1977, Oxygen and carbon isotope record through the Oligocene, DSDP Site 366, equatorial Atlantic, *in* Initial reports of the Deep Sea Drilling Project, Volume 41: Washington, D.C., U.S. Government Printing Office, p. 957–962.

Bolli, H. M., 1957, Planktonic foraminifera from the Oligocene-Miocene Cipero and Lengua Formations of Trinidad, B.W.I.: U.S. National Museum Bulletin, v. 215, p. 97–123.

Broecker, W. S., 1982, Ocean chemistry during glacial time: Geochimica et Cosmochimica Acta, v. 46, p. 1689–1705.

Bukry, D., 1972, Coccolith stratigraphy Leg 9, Deep Sea Drilling Project, *in* Initial reports of the Deep Sea Drilling Project, Volume 9: Washington, D.C., U.S. Government Printing Office, p. 817–832.

Crowley, T. J., and Matthews, R. K., 1983, Isotope-plankton comparisons in the late Quaternary core with a stable temperature history: Geology, v. 11, p. 295–278.

Dansgaard, W., 1964, Stable isotopes in precipitation: Tellus, v. 16, p. 436–468.

Douglas, R. G., and Savin, S. M., 1978, Oxygen isotopic evidence for the depth stratification of Tertiary and Cretaceous planktonic foraminifera: Marine Micropaleontology, v. 3, p. 175–196.

Duplessy, J-C., 1978, Isotope studies, *in* Gribbin, J., ed., Climatic change: Cambridge, Cambridge University Press, p. 46–67.

Graham, D. W., Corliss, B. H., Bender, M. L., and Keigwin, L. D., Jr., 1981, Carbon and oxygen isotopic disequilibria of recent deep-sea benthic foraminifera: Marine Micropaleontology, v. 6, p. 483–497.

Hays, J. D., Cook, H. E., et al., 1972, Initial reports of the Deep Sea Drilling Project, Volume 9: Washington, D.C., U.S. Government Printing Office, 1,205 p.

Keigwin, L. D., Jr., 1979, Late Cenozoic stable isotope stratigraphy and paleoceanography of DSDP sites from the east Equatorial and central North Pacific Ocean: Earth and Planetary Science Letters, v. 45, p. 361–382.

—— 1980, Palaeoceanographic change in the Pacific at the Eocene-Oligocene boundary: Nature, v. 287, p. 722–725.

Keller, G., 1983, Biochronology and paleoclimatic implications of Middle Eocene to Oligocene planktonic foraminiferal faunas: Marine Micropaleontology (in press).

Kennett, J. P., and Shackleton, N. J., 1976, Oxygen isotopic evidence for the development of the psychrosphere 38 m.y. ago: Nature, v. 260, p. 513–515.

Loutit, T. S., and Keigwin, L. D., Jr., 1982, Stable isotopic evidence for latest Miocene sea-level fall in the Mediterranean region: Nature, v. 300, p. 163–166.

Loutit, T. S., and Kennett, J. P., 1979, Application of carbon isotope stratigraphy to late Miocene shallow marine sediments, New Zealand: Science, v. 204, p. 1196–1199.

—— 1981, New Zealand and Australian Cenozoic sedimentary cycles and global sea-level changes: American Association of Petroleum Geologists Bulletin, v. 65, p. 1586–1601.

Loutit, T. S., Baum, G. R., and Wright, R. C., 1983, Eocene-Oligocene sea level changes as reflected in Alabama outcrop sections: American Association of Petroleum Geologists Bulletin, v. 67, p. 506.

Matthews, R. K., and Poore, R. Z., 1980, Tertiary $\delta^{18}O$ record and glacioeustatic sea-level fluctuations: Geology, v. 8, p. 501–504.

Miller, K. G., and Curry, W. B., 1982, Eocene to Oligocene benthic foraminiferal isotopic records in the Bay of Biscay: Nature, v. 96, p. 347–350.

Okada, H., and Bukry, D., 1980, Supplementary modification and introduction of code numbers to the low latitude coccolith biostratigraphic zonation (Bukry, 1973, 1975): Marine Micropaleontology, v. 5, p. 321–325.

Olsson, R. K., Miller, K. G., and Ungrady, T. E., 1980, Late Oligocene transgression of middle Atlantic coastal plain: Geology, v. 8, p. 549–554.

Pitman, W. C., III, 1978, Relationship between eustacy and stratigraphic sequences of passive margins: Geological Society of America Bulletin, v. 89, p. 1389–1403.

Poore, R. Z., and Matthews, R. K., 1984, Late Eocene–Oligocene oxygen and carbon isotope record from south Atlantic Ocean DSDP Site 522, *in* Initial reports of the Deep Sea Drilling Project, Volume 73: Washington, D.C., U.S. Government Printing Office (in press).

Quilty, P. G., 1977, Cenozoic sedimentation cycles in western Australia: Geology, v. 5, p. 336–340.

Savin, S. M., Douglas, R. G., and Stehli, F. G., 1975, Tertiary marine paleotemperatures: Geological Society of America Bulletin, v. 86, p. 1499–1510.

Shackleton, N. J., 1967, Oxygen isotope analyses and Pleistocene temperatures re-assessed: Nature, v. 215, p. 15–17.

—— 1977, Carbon-13 in *Uvigerina:* Tropical rainforest history and the equatorial Pacific carbonate dissolution cycles, *in* Andersen, N. R., and Malahoff, A., eds., The fate of fossil fuel CO_2 in the oceans: New York, Plenum, p. 401–427.

Shackleton, N. J., and Kennett, J. P., 1975, Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analyses in DSDP Sites 277, 279, 281, *in* Initial reports of the Deep Sea Drilling Project, Volume 29: Washington, D.C., U.S. Government Printing Office, p. 743–756.

Shackleton, N. J., and Opdyke, N. D., 1973, Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific Core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale: Quaternary Research, v. 3, p. 39–55.

Siesser, W. G., and Dingle, R. V., 1981, Tertiary sea-level movements around southern Africa: Journal of Geology, v. 89, p. 83–96.

Vail, P. R., and Hardenbol, J., 1979, Sea level changes during the Tertiary: Oceanus, v. 22, p. 71–80.

Vail, P. R., Mitchum, R. M., Jr., and Thompson, S., III, 1977, Global cycles of relative changes of sea level, *in* Seismic stratigraphy—applications to hydrocarbon exploration: American Association of Petroleum Geologists Memoir 26, p. 83–97

Van Andel, T. H., Health, G. R., and Moore, T. C. Jr., 1975, Cenozoic history and paleoceanography of the central equatorial Pacific Ocean: Geological Society of America Memoir 143, 134 p.

Vergnaud Grazzini, C., and Rabussier Lointier, D., 1980, Compositions isotopiques des foraminifères tertiaires en Atlantique equatorial (site 366 du DSDP): Revue de Géologie Dynamique et de Géographie Physique, v. 22, p. 63–74.

Vincent, E., Killingley, J. S., and Berger, W. H., 1980, The magnetic Epoch-6 carbon shift: A change in the oceans $^{13}C/^{12}C$ ratio 6.2 million years ago: Marine Micropaleontology, v. 5, p. 185–203.

Woodruff, F., Savin, S. M., and Douglas, R. G., 1981, Miocene stable isotope record: A detailed deep Pacific Ocean study and its paleoclimatic implications: Science, v. 212, p. 665–668.

ACKNOWLEDGMENTS

Supported by National Science Foundation Grant OCE-8008879. We thank C. Baker, B. H. Corliss, W. B. Curry, D. W. Graham, D. A. Johnson, and T. S. Loutit for helpful comments, E. Barron for reviewing the manuscript, and the National Science Foundation for providing samples through DSDP. Some of the ideas expressed in this paper evolved from discussions with W. B. Curry and T. S. Loutit. Woods Hole Oceanographic Institution Contribution No. 5395.

Manuscript received May 20, 1983
Revised manuscript received August 29, 1983
Manuscript accepted September 16, 1983