Stable isotopes in late middle Eocene to Oligocene foraminifera

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ABSTRACT

Oxygen and carbon isotope ratios in Eocene and Oligocene planktonic and benthic foraminifera have been investigated from Atlantic, Indian, and Pacific Ocean locations. The major changes in Eocene-Oligocene benthic foraminiferal oxygen isotopes were enrichment of up to 1‰ in ^{18}O associated with the middle/late Eocene boundary and the Eocene/Oligocene boundary at locations which range from 1- to 4-km paleodepth. Although the synchronous Eocene-Oligocene ^{18}O enrichment began in the latest Eocene, most of the change occurred in the earliest Oligocene. The earliest Oligocene enrichment in ^{18}O is always larger in benthic foraminifera than in surface-dwelling planktonic foraminifera, a condition that indicates a combination of deep-water cooling and increased ice volume. Planktonic foraminiferal $\delta^{18}O$ does not increase across the middle/late Eocene boundary at our one site with the most complete record (Deep Sea Drilling Project Site 363, Walvis Ridge). This pattern suggests that benthic foraminiferal $\delta^{18}O$ increased 40 m.y. ago because of increased density of deep waters, probably as a result of cooling, although glaciation cannot be ruled out without more data.

Stable isotope data are averaged for late Eocene and earliest Oligocene time intervals to evaluate paleoceanographic change. Average $\delta^{18}O$ of benthic foraminifera increased by 0.64‰ from the late Eocene to the early Oligocene $\delta^{18}O$ maximum, whereas the average increase for planktonic foraminifera was 0.52‰. This similarity suggests that the Eocene/Oligocene boundary $\delta^{18}O$ increase was caused primarily by increased continental glaciation, coupled with deep sea cooling by as much as 2 °C at some sites. Average $\delta^{18}O$ of surface-dwelling planktonic foraminifera from 14 upper Eocene and 17 lower Oligocene locations, when plotted versus paleo-latitude, reveals no change in the latitudinal $\delta^{18}O$ gradient. The Oligocene data are offset by ~0.45‰, also believed to reflect increased continental glaciation. At present, there are too few deep sea sequences from high latitude locations to resolve an increase in the oceanic temperature gradient from Eocene to Oligocene time using oxygen isotopes.

INTRODUCTION

The history of Cenozoic stable isotopic change in benthic foraminiferal calcite is marked by a 4‰ to 5‰ increase in the ratio of ^{18}O to ^{16}O. This over-all change is thought to reflect the evolution of climate through growth of continental ice and cooling of the deep ocean (Emiliani, 1954; Savin and others, 1975; Shackleton and Kennett, 1975), although the timing and relative importance of these effects have been questioned (Matthews and Poore, 1980). The oxygen isotopic record of surface-dwelling

planktonic foraminifera closely parallels that of benthic foraminifera at high-latitude locations (Shackleton and Kennett, 1975) but shows less change at lower latitudes (Savin and others, 1975), indicating that there has been an increase in the planetary temperature gradient through the Cenozoic which was reflected in greater cooling of high-latitude than of low-latitude surface waters.

The Cenozoic record of foraminiferal carbon isotopic composition is variable within about 2‰ limits except during the Paleocene when $\delta^{13}C$ increased by ~2‰ (Shackleton and others, 1984). Carbon isotopic variation on a million-year (m.y.) time scale may reflect variability in the burial of organic matter. Shackleton (1985) has investigated this effect in the Cenozoic by means of the $\delta^{13}C$ record in bulk (mostly coccolith) carbonate. He concluded that the long-term decrease in $\delta^{13}C$, especially since the middle Miocene, may have been the result of a decrease in size of the global organic carbon reservoir. Higher frequency (10^4-10^5 yr) variability in $\delta^{13}C$ may reflect transfer of organic carbon between the open ocean and continents (Shackleton, 1977), between the open ocean and continental shelves and estuaries (Broecker, 1982), or between deep basins of the ocean (Kroopnick, 1974).

Close inspection of the long-term record of oxygen isotopic change in Cenozoic carbonates reveals that much of the over-all change occurred abruptly during a few intervals (Savin and others, 1975; Shackleton and Kennett, 1985). The most well defined of these times appear to be near the Eocene/Oligocene boundary, within the middle Miocene, and within the Pliocene and Pleistocene. This paper focuses on the earliest of these changes, from middle Eocene to early Oligocene time. This interval is of particular interest because of various geological events that occurred, including a turnover in oceanic faunas and floras (Corliss and others, 1984). We have chosen for detailed study reasonably complete deep-sea sequences (Fig. 1) from the Atlantic Ocean (DSDP 19, DSDP 363, and Eureka 67–128), the Indian Ocean (DSDP 219 and 253), and the Pacific Ocean (DSDP 77, 277, and 292). The objectives of this study are (1) to establish the timing of Eocene-Oligocene stable isotopic change at these locations, (2) to determine the nature of Eocene-Oligocene paleoceanographic change, and (3) to determine planetary temperature gradient changes from late Eocene to early Oligocene time.

PREVIOUS WORK

Shackleton and Kennett (1975) studied the entire Cenozoic with wide sampling intervals and later conducted a more detailed study of the latest Eocene to earliest Oligocene at DSDP 277 (Kennett and Shackleton, 1976). The more closely spaced results showed that the ^{18}O enrichment occurred within lowermost Oligocene sediment and that it occurred in as little as 0.1 m.y. Boersma and Shackleton (1978) were the first to study in
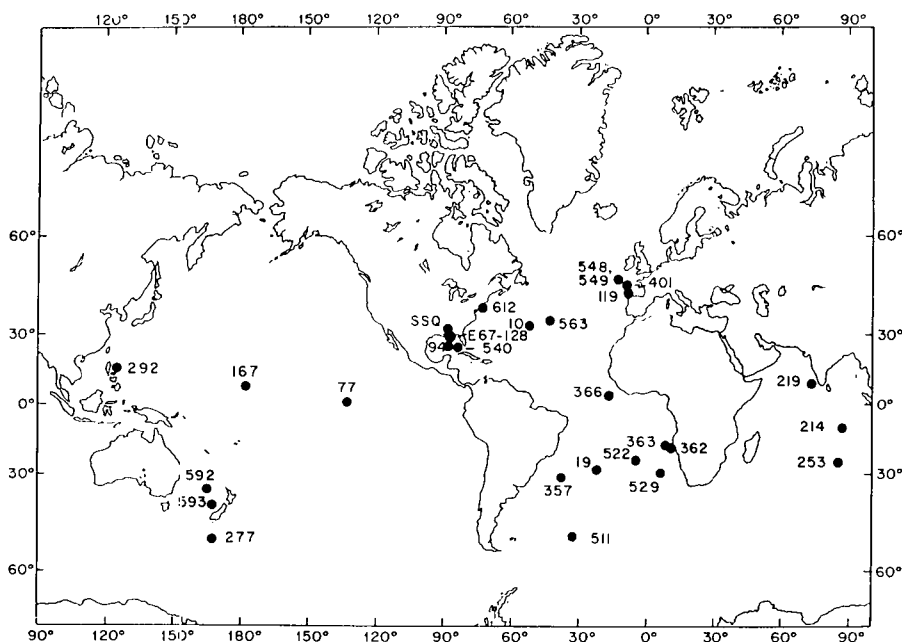

Figure 1. Site locations discussed in this study.

detail Eocene-Oligocene stable isotopic change in an Atlantic core (Site 357, Rio Grande Rise). Site 357 is marked by several unconformities, and analysis of one sample placed the major Eocene/Oligocene ^{18}O enrichment within the late Eocene. Because that sample came from foraminiferal Zone P17 and because P17 is now thought to straddle the Eocene/Oligocene boundary (Hardenbol and Berggren, 1978), the exact age of that sample is uncertain. Vergnaud-Grazzini and others (1978) reported isotopic results from three locations in the northeast Atlantic Ocean (DSDP 398, 400A, and 401), but diagenesis lowered their $\delta^{18}O$ values. Miller and Curry (1982) subsequently analyzed samples from less deeply buried sites in the same region and obtained results more in accordance with those of other studies. Keigwin (1980) re-examined the isotopic record at DSDP 277 and at DSDP 292, using monospecific benthic samples to rule out any change in the earlier record of Kennett and Shackleton (1976) which might have been due to analysis of mixed benthic foraminifera. Those results, spanning 70° of paleo-latitude, showed that the ^{18}O enrichment began at the Eocene/Oligocene boundary at each location and occurred in <0.9 m.y. The indication is that this change was more the result of a cooling than that of an ice volume increase. Many subsequent studies have revealed essentially the same patterns of isotope enrichment (for example, Muza and others, 1983; Shackleton and others, 1984; Oberhansli and others, 1984; Murphy and Kennett, in press).

METHODS

Analytical procedures were similar to those described in the report of Keigwin (1979). Our reference gas is calibrated to the standard PDB through repeated reaction and analysis of the powder $CaCO_3$ standard NBS-20 in the same manner as unknowns except that the standard is not roasted in vacuo. The $\delta^{18}O$ and $\delta^{13}C$ of NBS-20 were taken to be $-4.18\%o$ and $-1.06\%o$, respectively. Reproducibility (1σ) of standard analyses using these procedures is better than $0.10\%o$ for both carbon and oxygen isotopes. Foraminifera in this study were rarely abundant enough for replicate analyses.

At most locations, we have chosen for analysis the benthic foraminiferal species Oridorsalis tener, Globocassidulina subglobosa, and species of

Cibicidoides. Although we consider O. tener less than an ideal species for the reasons cited by Belanger and others (1981), Savin and others (1981), and Keigwin (1982), it is cosmopolitan, wide ranging, well preserved, and frequently the only benthic species abundant enough for analysis. Wherever possible, we analyzed Cibididoides, assuming the Paleogene species fractionate stable isotopes in a manner similar to modern Cibicidoides (Belanger and others, 1981; Graham and others, 1981).

For planktonic foraminifera, we are mostly concerned with the record of surface-dwelling species. At DSDP Sites 292 (Fig. 2) and 77B (Fig. 3), we analyzed several planktonic foraminiferal taxa and plotted oxygen- versus carbon-isotope results to help determine which species lived closer to the sea surface. Surface-dwelling species should have both lower $\delta^{18}O$, reflecting carbonate precipitation at higher temperatures, and greater $\delta^{13}C$ than more deeply dwelling planktonics. Although we cannot constrain species-dependent disequilibrium effects, our results reveal, in general, that Pseudohastigerina, Chiloguembelina, and Globigerina ampliapertura are surface-dwelling taxa. At Site 77B, Subbotina linaperta is seen to be an intermediate-depth dweller; at both Sites 77B and 292, species of Catapsydrax lived so deep that they have isotope values close to those of benthic foraminifera. These observations are supported by additional analyses of planktonic foraminiferal species at other sites (see Appendix 1)[1] and are consistent with the results of Poore and Matthews (1984b).

EFFECTS OF DIAGENESIS

The diagenetic change from ooze to chalk to limestone generally occurs as a function both of depth of burial and time, although this relationship is complicated by paleoceanographic conditions specific to each location (Schlanger and Douglas, 1974). Basically three processes affect the postdepositional oxygen isotopic composition of foraminifera: (1) recrystallization at higher in situ temperature, (2) recrystallization in equilibrium with pore waters enriched in ^{16}O due to basalt and clay mineral alteration (Anderson and others, 1976), and (3) recrystallization

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[1]Appendices 1 and 2 may be secured free of charge by requesting Supplementary Data 86-05 from the GSA Documents Secretary.
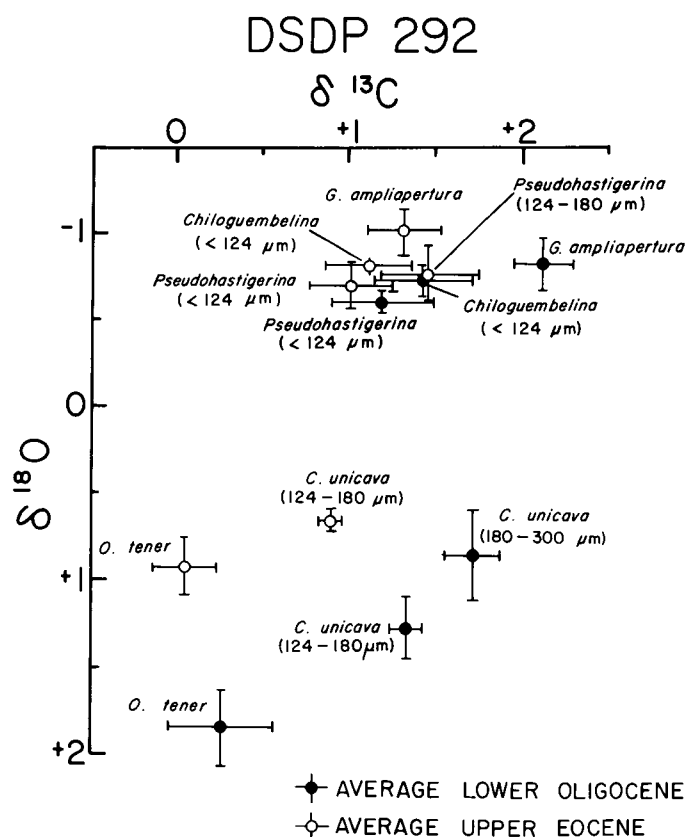
Figure 2. Average upper Eocene and lower Oligocene stable isotope results at Site 292. Surface-dwelling planktonic foraminifera are considered to be those with low $\delta^{18}O$ and high $\delta^{13}C$ values.
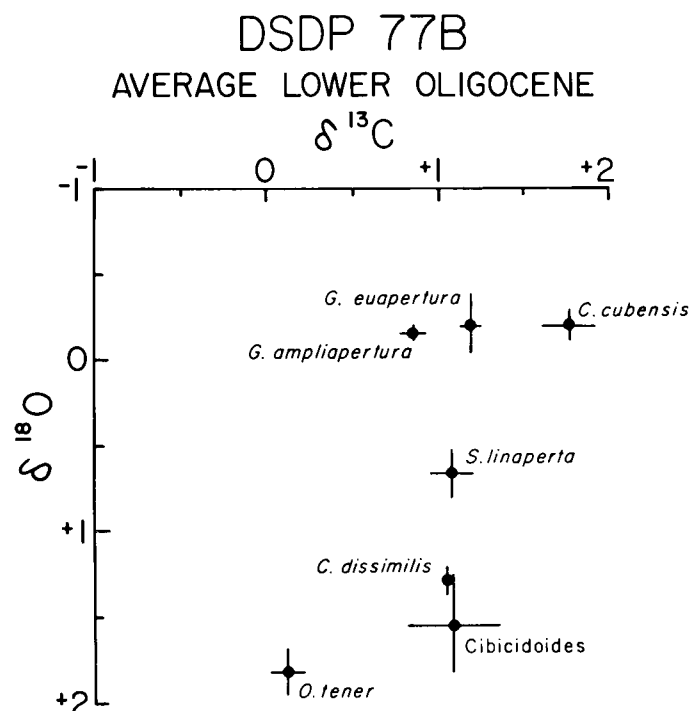

Figure 3. Average lower Oligocene stable isotope results at Site 77B. Surface-dwelling planktonic foraminifera are considered to be those with low $\delta^{18}O$ and high $\delta^{13}C$ values.

in a closed system (Killingley, 1983). The Killingley model for recrystallization proposes that benthic foraminiferal $\delta^{18}O$ decreases toward planktonic values (approximately the value of bulk carbonate) as diagenesis progresses. It is thus possible that the Paleogene to Neogene trend of diverging benthic and planktonic foraminiferal $\delta^{18}O$ values could be a diagenetic artifact.

Distinguishing between these processes is not always possible. For example, in some regions there may be both high heat flow and basalt–pore-water interaction as found by Mottl and others (1982) at Site 504 in the Panama basin. A similar situation can probably account for the diagenesis and lowered $\delta^{18}O$ observed elsewhere in the Panama basin (Keigwin, 1979). Where overburden exceeds several hundred metres, a small percentage of recrystallization combined with a geothermal temperature gradient of ~35 °C km^{-1} is sufficient to cause significant lowering of $\delta^{18}O$ values. This is probably the case in eastern North Atlantic DSDP Sites 400A and 398 (~550 to 600 m; Vergnaud-Grazzini and others, 1978), which were found to have $\delta^{18}O$ values of benthic foraminifera as much as 2‰ lower than shallower buried sediment of the same age (~100–400 m; Miller and Curry, 1982).

A comparison of $\delta^{18}O$ results from Sites 362A (Fig. 4) and 363 (Fig. 5) indicates that CaCO$_3$ is altered at Site 362A. Site 362A is located very close to Site 363 on the Walvis Ridge but is in shallower water. At about 1,000-m burial depth, the Eocene/Oligocene sediments of Site 362A are nearly twice as deep as those recovered at Site 593 (546 m), the next deepest of sites studied here. The benthic foraminiferal isotope record

contains all of the features found at Site 363, notably ^{18}O enrichment associated with the middle/upper Eocene boundary and the Eocene/Oligocene boundary and within the middle of the Oligocene. Samples most enriched in ^{18}O come from the middle of the Oligocene, perhaps time-equivalent with the $\delta^{18}O$ maximum reported from the Pacific by Keigwin and Keller (1984) and Miller and Thomas (1985). Site 362A differs from all other sites in this study in its very low $\delta^{18}O$ values, which are about 1‰ less than equivalent values at Site 363. This condition most likely results from carbonate diagenesis, first noted by shipboard scientists (Bolli, Ryan, and others, 1978). Even when altered, however, $\delta^{18}O$ still may be useful for isotope stratigraphy, as the altered Bay of Biscay and Walvis Ridge sites show the same trends as the well-preserved sites.

Carbon isotopes at most deep-sea locations do not appear to change significantly even when there is severe alteration in foraminiferal $\delta^{18}O$. There may be two reasons for this: (1) the temperature coefficient for carbon isotope fractionation during calcium carbonate precipitation is about one-tenth that for oxygen isotopes; (2) in the absence of significant amounts of organic matter, the carbon reservoir in pore waters is much smaller than that in carbonates, and thus the carbon-isotope composition of recrystallized carbonate will largely reflect the composition of the primary calcite. This may account for the observation at Site 362A that although $\delta^{18}O$ in the benthic foraminifera Cibicidoides is 1‰ lower due to depth of burial than at Site 363, the $\delta^{13}C$ values are equivalent. Foraminiferal $\delta^{13}C$ may still be useful, therefore, in paleoceanography, even if there has been a diagenetic effect on $\delta^{18}O$.

Most of our samples lack obvious evidence of recrystallization, and we assume that our $\delta^{18}O$ data reflect primary climatic and oceanographic information: the temperature and composition of sea water. Using the SEM, we examined specimens of Cibicidoides and O. tener from Sites 253
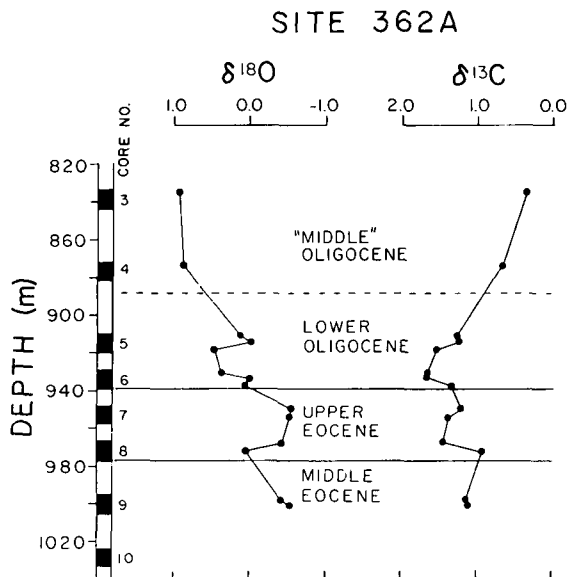
Figure 4. Stable isotope results on *Cibicidoides* at Site 362A. Burial depth is considerably greater and oxygen-isotope ratios at this site are about 1‰ lower than nearby Site 363 (Fig. 5). Lowered $\delta^{18}O$ is evidence of calcite diagenesis.
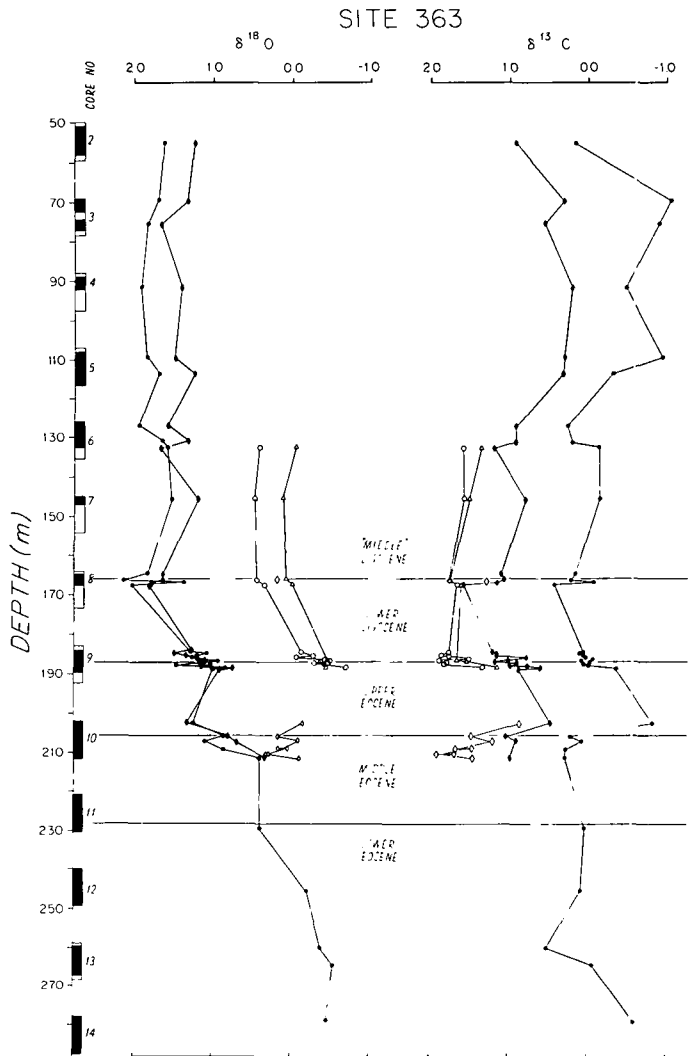

Figure 5. Stable isotope results at DSDP Site 363. Solid data points are benthic foraminifera (dots = *O. tener*; diamonds = *Cibicidoides*). Open points are planktonic foraminifera (circles = *G. ampliapertura*; triangles = *Chiloguembelina*; diamonds = *Pseudohastigerina*). Sediment recovery is shown in black along vertical axis. Note similar patterns of benthic and planktonic $\delta^{18}O$ at the Eocene/Oligocene boundary, suggesting influence of ice volume. At the middle/upper Eocene boundary, however, the lack of covariance suggests greater influence of cooling deep water. Divergence of $\delta^{18}O$ values for *O. tener* and *Cibicidoides* in Oligocene time indicate changing "vital effects" of *O. tener*.

and 363 because of a curious divergence of $\delta^{18}O$ values for these species from the Eocene to Oligocene, which might be evidence of diagenesis (Figs. 5 and 6). Careful examination of inner and outer chamber walls revealed no obvious recrystallization. We did, however, find surface features indicative of $CaCO_3$ dissolution (Corliss and Honjo, 1981) and occasional cemented coccoliths which were not dislodged ultrasonically. Sr-isotope ratios in pore water at Site 253 also indicated carbonate recrystallization reactions (Elderfield and Gieskes, 1982). Sr/Ca results on seven samples of late Eocene *Chiloguembelina* spp. from Site 253 were provided by M. Delaney (then at Massachusetts Institute of Technology), who used graphite furnace atomic absorption. The samples were extremely small and close to blank levels and detection limits, but five analyses were lower than results on other samples that age (Graham and others, 1982), suggesting the possibility of some recrystallization at that site. Generally consistent results at locations from diverse geographic and oceanographic settings, however, argue against widespread alteration of our Eocene and Oligocene stable isotope values.

STUDY LOCATIONS

Assignment of epoch and sub-epoch boundaries comes from the biostratigraphic studies of Keller (1982/1983) for Sites 363, 253, 219, and Keller (1985, personal commun.) for E67-128; Aubry (unpub. data) for Site 19; and Bolli and others (1978) for Site 362A. We use the term "middle" Oligocene in an informal sense to refer to the *G. ampliapertura* and *G. opima* Zones.

Site 363, in the easternmost Atlantic on the aseismic Walvis Ridge, has the longest record in terms of time and length of sediment core of all of our sites (Fig. 5). Unfortunately, it was only "spot-cored." The rapid increase in $\delta^{18}O$ in two species of benthic foraminifera associated with the middle/late Eocene boundary is not evident in the planktonic foraminifera. This observation indicates either that bottom waters at this location

cooled or the unlikely possibility that there was a synchronous increase in continental ice and a 4 °C warming of surface waters. This pattern must be verified elsewhere. The $\delta^{18}O$ difference between species of benthic foraminifera and between planktonic species increases in Oligocene samples. Although this increase might possibly be a diagenetic artifact, severe diagenesis would be expected to drive benthic $\delta^{18}O$ values closer to planktonic values (Killingley, 1983).

Site 19, the deepest of our sequences (4,685-m water depth), was cored on the west flank of the Mid-Atlantic Ridge (Fig. 7). Planktonic foraminifera suffered selective solution, and only the most robust, inferred
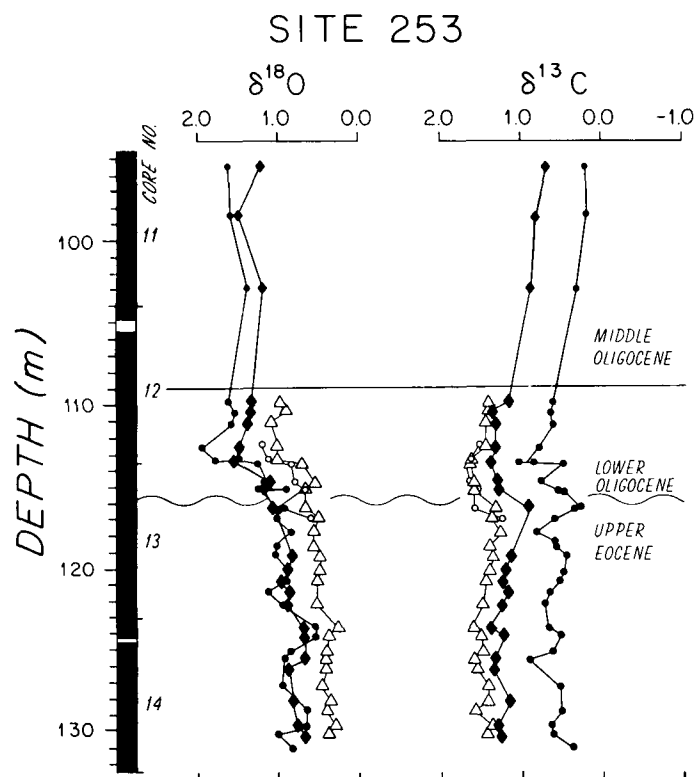
Figure 6. Stable isotope results at Site 253 (symbols as in Fig. 5). As at Site 363, note increasing difference between *C. ungerianus* and *O. tener* $\delta^{18}O$ data in Oligocene time.
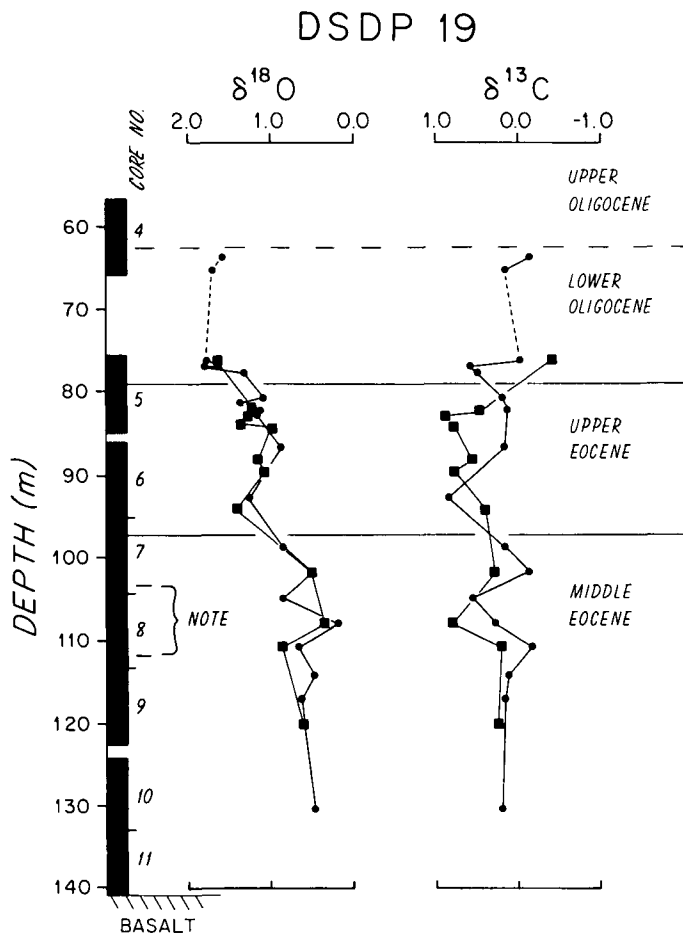
Figure 7. Stable isotope results on benthic foraminifera at Site 19 (symbols as in Fig. 5). Studies of planktonic foraminifera and nannofossils suggest a possibly repeated sequence from the bottom of core 7 to the bottom of core 8 (Maxwell, von Herzen, and others, 1970). Surface-dwelling planktonic foraminifera are absent due to dissolution. Also note increases in $\delta^{18}O$ associated with the middle/late Eocene and Eocene/Oligocene boundaries.

deep-dwellers are preserved. Benthic foraminifera show the trend of ^{18}O enrichment at the middle/upper Eocene boundary and from the upper Eocene to lower Oligocene.

In the Gulf of Mexico region, we sampled a time series at Eureka 67-128 and St. Stephens Quarry in Alabama (Mancini, 1979). The benthic foraminiferal fauna at E67-128 differs from that at our other sites, and so we chose for analysis *Bulimina alazanensis* because it was abundant and long ranging. This species shows the typical enrichment in ^{18}O of nearly $1^{0}/_{00}$ associated with the Eocene/Oligocene boundary, but planktonic specimens do not follow this trend (see Appendix 1), perhaps because of reworking (Poore and Matthews, 1984b). Of our analyses at St. Stephens Quarry (Appendix 1), we have considered only the samples from the lower Oligocene Red Bluff Clay (Mancini, 1979) because underlying and overlying samples show evidence of calcite overgrowths and recrystallization.

In addition to Site 253 (Fig. 6) on the Ninetyeast Ridge in the Indian Ocean, we examined the isotope record at Site 219 (Fig. 8) on the Laccadive-Chagos Ridge in the Arabian Sea. Results at the latter site differ from our other site results in two respects. (1) A small enrichment in ^{18}O in benthic foraminifera is found below the middle/upper Eocene boundary, whereas at Sites 19 and 363 a larger change occurs at, or across, the boundary. Perhaps the larger middle/late Eocene boundary $\delta^{18}O$ increase seen at Sites 19 and 363 is not evident because of an unconformity (Keller, 1982/83). (2) Although maximum benthic foraminiferal $\delta^{18}O$ values are found in lowest Oligocene sediment, the change is not as large and abrupt as at other sites. This probably is the result of severe coring disturbance in the upper part of core 16.

Since the publication of earlier results (Kennett and Shackleton, 1976; Keigwin, 1980), we have completed additional stable isotopic work at DSDP Sites 277 and 292 in the Pacific Ocean. General trends at these sites remain unchanged. At Site 277, we analyzed the benthic foraminifera *G. subglobosa* and the planktonic genus *Chiloguembelina*, which has lower $\delta^{18}O$ values than *Globigerina angiporoides*. Murphy and Kennett (in press) have recently completed a study of Eocene/Oligocene stable isotopes from southwest Pacific DSDP locations, which we have used in guiding our sampling of Sites 592 and 593 for paleoceanographic reconstructions.

Figure 9, which summarizes benthic foraminiferal isotope results versus age from our sites with the most complete time series, reveals maxima in $\delta^{18}O$ at about 40 Ma, 35-36 Ma, and 29-30 Ma. The maximum near 40 Ma (the middle/late Eocene boundary), which has not been reported in previous studies, is not well defined in Figure 7. This may result from small errors in age assignment. The 35-36 Ma (lower Oligocene) maximum has been observed previously by many workers and the 29-30 Ma (middle Oligocene) maximum found by Keigwin and Keller (1984) has been
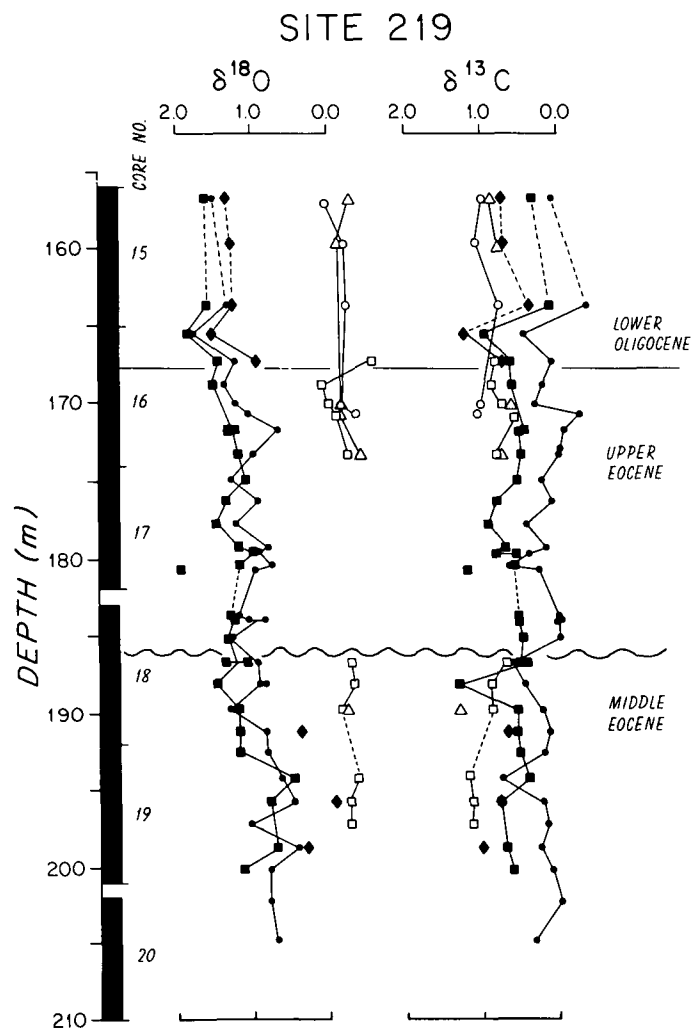
SITE 219

$\delta^{18}O$ $\delta^{13}C$

2.0 1.0 0.0   2.0 1.0 0.0

CORE NO.

DEPTH (m)

160 — 15

170 — 16

180 — 17

190 — 18

200 — 19

210 — 20

LOWER OLIGOCENE

UPPER EOCENE

MIDDLE EOCENE

Figure 8. Stable isotope results at Site 219. Symbols as in Figure 5, except *Pseudohastigerina* data are shown as open squares. The $\delta^{18}O$ increases seen at other sites at the Eocene/Oligocene boundary and at the middle/late Eocene boundary may not be evident at Site 219 due to severe coring disturbance in the top of core 16 and to a hiatus in core 18 (Keller, 1982/1983).
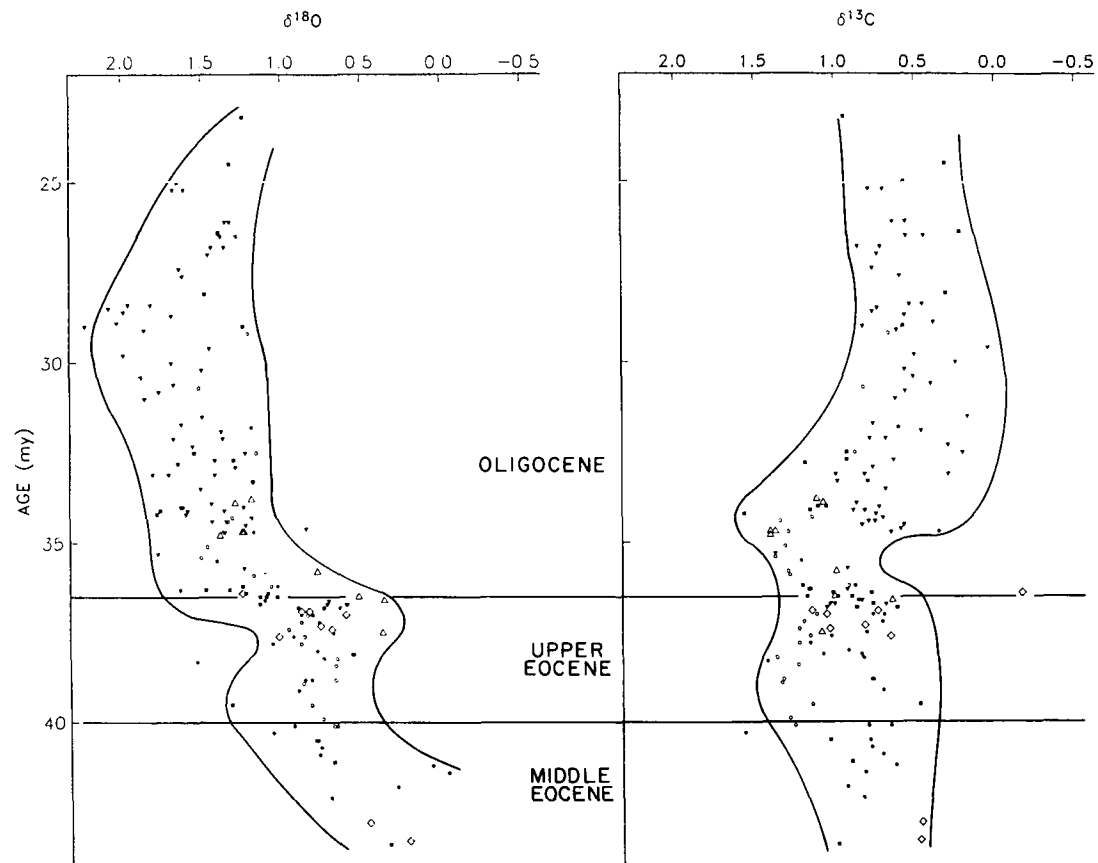
$\delta^{18}O$

2.0 1.5 1.0 0.5 0.0 -0.5

$\delta^{13}C$

2.0 1.5 1.0 0.5 0.0 -0.5

AGE (my)

25

30

35

40

OLIGOCENE

UPPER EOCENE

MIDDLE EOCENE

Figure 9. Oxygen- and carbon-isotope results versus age in *Cibicidoides*, or species "corrected" to *Cibicidoides* for Sites 77B (solid triangles), 363 (squares), 19 (diamonds), 219 (solid circles), 253 (open circles), and 277 (open triangles). Oxygen isotopes show enrichment at ~40 Ma, 36 Ma, and 29 Ma; carbon isotopes reveal a maximum at 36 Ma.

TABLE 1. PAIRED ANALYSES OF BENTHIC FORAMINIFERA

Species pairs	No. of pairs	δ	Mean ‰	1σ
C. ungerianus-O. tener (all data)	76	^{18}O	-0.17	0.23
		^{13}C	+0.84	0.27
Oligocene C. ungerianus-O. tener*	45	^{18}O	-0.21	0.24
		^{13}C	0.90	0.29
Eocene C. ungerianus-O. tener*	31	^{18}O	-0.15	0.20
		^{13}C	0.78	0.21
C. ungerianus-G. subglobosa*	12	^{18}O	-0.39	0.25
		^{13}C	+0.27	0.14
C. ungerianus-Bulimina sp.*	5	^{18}O	-0.32	0.11
		^{13}C	+1.01	0.41
C. ungerianus-Uvigerina	10	^{18}O	-0.36	0.17
	11	^{13}C	0.70	0.41

*Corrections used in paleoceanographic reconstructions (Figs. 9 and 10).

confirmed by Miller and Thomas (1985). For stratigraphic purposes, it is unimportant whether the pattern of ^{18}O enrichment reflects cooling of bottom waters or growth of continental ice; it appears to be a general feature of deep-sea isotope records. For paleoceanography, however, the interpretation of δ^{18}O has considerable importance. Oxygen-isotope evidence for glaciation as early as 35–36 Ma has been reported recently by several workers (Keigwin, 1980; Keigwin and Keller, 1984; Miller and Fairbanks, 1983; Poore and Matthews, 1984a; Shackleton and others, 1984; Miller and Thomas, 1985). Basically, most workers agree that δ^{18}O values >2.0‰ on *Cibicidoides* from deep-sea locations suggest the presence of continental ice, because deep-sea temperatures would have been colder than today if no ice was present.

Carbon-isotope results from our cores display great variability; the most significant event is a maximum in δ^{13}C at ~35 Ma (Fig. 7). This event has also been reported by Miller and Fairbanks (1983, 1985), whose more complete Oligocene time series reveals cyclic changes in δ^{13}C.

PALEOCEANOGRAPHIC RECONSTRUCTIONS

Shackleton and Boersma (1981) presented the only previous reconstruction of Paleogene climatic and oceanographic conditions based on stable isotope data. Their major conclusion was that the Eocene world had less than one-half the latitudinal temperature gradient of the present world, so that ocean currents must have been more important in the poleward transport of heat than they are at present.

We attempt to establish how the ocean changed from late Eocene to Oligocene time using isotope data. An underlying assumption is that changes from the Eocene to the Oligocene were greater than the variability within either time-stratigraphic interval so that late Eocene data can be "pooled" and compared with "pooled" early Oligocene data. At most locations we averaged Eocene data from the *G. cerroazulensis* Zone. In synthesizing Oligocene data, we disregarded those samples before the early Oligocene maximum in benthic foraminiferal δ^{18}O in order to emphasize the contrast between the Eocene and Oligocene. The presence or absence of continental ice will not affect our reconstructions, because the sea-water composition effect will be the same for all sites within each time-stratigraphic interval. To our primary data base of sites where we have a time series (Figs. 4–8), we have added data from studies of others and from additional sites where we have more limited data (Appendix 1).

Paired Analyses of Benthic Foraminifera

The same species of benthic foraminifera were not abundant enough for isotopic analysis at each location, and paired analyses of species were performed in order to determine "correction factors." Such corrections are commonly used to account for non-equilibrium fractionation ("vital effects") between several species (Boersma and Shackleton, 1977; Graham and others, 1981; Savin and others, 1981; Shackleton and others, 1984). It is assumed that correction factors between species are invariant through time (hence, they are genetically controlled), although this has been demonstrated for only a few species pairs in the Neogene (Savin and others, 1981; Graham and others, 1981). Some species, notably *Oridorsalis* spp., have been found to have a ^{13}C/^{12}C ratio, which varies compared to *Cibicidoides* spp. (Belanger and others, 1981; Savin and others, 1981; Keigwin, 1982).

All benthic foraminiferal stable isotope data were "corrected" to *Cibicidoides,* on the basis of the differences between paired analyses of species summarized in Table 1. Although (as pointed out by Savin and others, 1981), using "correction factors" is certain to introduce artifacts of a few tenths per mil into benthic foraminiferal data sets, this technique is the only hope for comparing data from sites where *Cibicidoides* are absent.

It is evident from Figures 5 and 6 that the δ^{18}O difference between *Cibicidoides* and *O. tener* has changed through time, although considering all pairs from all sites, the difference is insignificant (Table 1). Nevertheless, we used separate Eocene and Oligocene corrections for these pairs. The origin of the changing $\Delta\delta^{18}$O is unknown. Diagenesis seems unlikely for the reasons discussed earlier, although it cannot be completely ruled out. Regardless of what caused the $\Delta\delta^{18}$O to change from Eocene to Oligocene time, the Paleogene difference is substantially less than differences of about –0.60‰ to –0.70‰ observed in pairs of Neogene *Cibicidoides* and *Oridorsalis* (Savin and others, 1981; Graham and others, 1981). It should also be noted that our δ^{18}O difference between *Uvigerina* and *Cibicidoides* (–0.36‰; Table 1) is significantly less than the difference (–0.60‰ to –0.70‰) widely quoted for Neogene pairs of these genera (Shackleton and Opdyke, 1973; Woodruff and others, 1980; Savin and others, 1981; Belanger and others, 1981; Graham and others, 1981).

Backtracking

All sites are backtracked for latitude and water depth to their positions at the Eocene/Oligocene boundary (Table 2; Appendix 2).[2] Paleolatitude determinations are taken from the following maps: Pacific Ocean, from van Andel and others (1975) and Molnar and others (1975); Atlantic Ocean, from Sclater and others (1977b); and Indian Ocean, from Sclater and others (1977a).

The calculation of most paleodepths in Appendix 2 comes from the procedure of Tucholke and Vogt (1979), using their subsidence curve. For the Pacific Ocean and Indian Ocean, the subsidence curve of van Andel and Bukry (1973) was used. Essentially, basement paleodepth was calculated for 36 Ma and then corrected for sediment underlying the Eocene/Oligocene boundary. Such calculations probably have error limits of a few hundred metres, provided that the sites come from "normal" ocean crust (Sclater and McKenzie, 1973). Many sites are from areas of anomalously shallow water depth, chosen so that the sea floor would be above the calcite compensation depth at the time of interest. For aseismic ridges, Detrick and others (1977) showed that subsidence occurs at rates similar to normal oceanic crust. It is not clear, however, how long sites on other features such as plateaus (Sites 277, 592, 593) or rises (Site 292) have been shallow.

The greatest uncertainty in the paleodepth determination probably comes from uncertainty in basement age and depth. In young oceanic crust, small errors in age can account for significant differences in paleodepth, whereas older crust may not have subsided appreciably in the last 37 m.y. At most of our sites with uncertain basement ages, we have assumed minimum values. When considerable uncertainty prevented calculation of Eocene/Oligocene boundary sea-floor paleodepth, the present depth was assumed (Table 2).

Vertical Reconstructions

Sites in this study ranged in paleodepth from ~500 to 4,000 m. Many chemical and physical gradients that might be recorded in the stable isotope ratios of benthic foraminifera are found throughout this depth range in the modern ocean. Trends in the δ^{13}C of ΣCO_2, for example, accurately reflect trends in dissolved O_2 in the water column (Kroopnick and others, 1972). As the present deep interocean difference in δ^{13}C of ΣCO_2 reflects the present pattern of deep-sea circulation (Kroopnick, 1974), and as this pattern is also seen in the δ^{13}C of the benthic foraminifera *Cibicidoides* (Graham and others, 1981; Belanger and others, 1981), studies of carbon

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[2]See footnote 1.

TABLE 2. SUMMARY OF STABLE ISOTOPE RESULTS, PALEODEPTH, AND PALEOLATITUDE OF SITES CONSIDERED IN THIS STUDY

Site	Lat. at 37 Ma	Depth (m)* at 37 Ma	Reference	Species	Late Eocene δ¹⁸O	δ¹³C	Early Oligocene δ¹⁸O	δ¹³
10		4,227		Cibicidoides			1.58	1.53
19		3,730		O. tener			1.22	1.07
				G. subglobosa	0.82	0.91		
77B	8°S	2,651		Cibicidoides	0.65	0.93	1.55	1.08
				C. cubensis			-0.18	0.78
94	20°N	1,793†		Cibicidoides	0.87	1.16	1.48	1.20
				P. micra	-0.84	0.87		
				C. cubensis			-0.41	1.07
119		3,700§	Miller and Curry (1982)	Cibicidoides			1.58	0.57
167	1°S	3,363		Cibicidoides	0.67	0.81		
				C. cubensis			-0.30	1.55
214		859		Cibicidoides	0.69	1.20	1.17	1.16
219	2°S	1,168		Cibicidoides			1.25	0.79
				G. subglobosa	0.84	0.87		
				G. ampliapertura	-0.30	1.00		
				Chiloguembelina spp.			0.21	0.81
253	40°S	606		Cibicidoides	0.82	1.22	1.34	1.07
				Chiloguembelina spp.	0.45	1.44	0.99	1.41
277	59°S	1,232†		G. subglobosa	0.42	0.92	1.27	1.28
				Chiloguembelina spp.	-0.13	1.56	0.39	1.62
292	6°N	1,068		O. tener	0.77	0.91	1.63	1.11
				G. ampliapertura	-1.01	1.32	-0.81	2.12
357	39°S	1,500		Cibicidoides			1.21	0.46
				C. cubensis			-0.04	0.55
363	30°S	2,216		Cibicidoides	1.04	0.77	1.49	0.83
				G. ampliapertura	-0.48	1.65		
				Chiloguembelina spp.			0.03	1.55
366	4°S	2,601		Cibicidoides	0.60	0.60	1.50	1.26
				C. cubensis			-1.13	1.63
			Poore and Matthews (1984b)	G. pomeroli	-1.50	1.04		
401	40°N	2,495†	Miller and Curry (1982)	Cibicidoides	0.95	0.78		
				C. cubensis, Pseudohastigerina	0.20	1.11		
511	59°S		Muza and others (1983)	G. angiporoides, G. aff.linaperta	0.40	0.58		
522	35°S	2,693	Poore and Matthews (1984a)	Cibicidoides	1.23	0.99	1.97	1.23
			Poore and Matthews (1984a)	G. ampliapertura			1.21	1.79
			Poore and Matthews (1984b)	H. alabamensis	0.25	1.86		
529	40°S	2,362	Shackleton and others (1984)	Cibicidoides			1.71	1.06
			Shackleton and others (1984)	G. cerroazulensis, T. increbescens, G. index	0.30	2.15		
540	20°N	2,926†	Belanger and Matthews (1984)	Cibicidoides	0.94	1.34	1.28	0.93
				C. cubensis			-0.29	1.10
548A	40°N	969		Cibicidoides			0.97	0.97
				C. cubensis			0.18	1.00
549		2,386	Miller and others (1985)	Cibicidoides	0.70	1.10	1.28	1.30
563		2,200	Miller and Fairbanks (1985)	Cibicidoides			1.86	1.24
592	43.5°S	1,098†		Cibicidoides	0.13	1.25	1.02	2.03
				Chiloguembelina spp.	-0.06	1.64	0.44	2.00
593	47.5°S	1,068†		Cibicidoides	0.19	1.31	0.74	1.86
				Chiloguembelina spp.	-0.22	1.64	0.25	2.00
612	33°N	1,386†		Cibicidoides	0.51	-0.05		
				P. micra	-0.75	-0.36		
				P. gemma			-0.19	-1.29
E67-128	25°N	1,494†		B. alazanensis	-0.04	0.33	0.70	0.39
				G. ampliapertura	-1.39	1.08		
				Chiloguembelina spp.			-1.34	-0.66
SSQ	28°N			Pseudohastigerina spp.			-1.54	-0.37

*From Appendix 2 (GSA Data Repository).
†For these locations, present-day water depth is assumed.
§Present water depth corrected by change in depth in 37 m.y.

isotopes in fossil foraminifera can provide information on the evolution of deep-sea circulation. For the late Quaternary, this has been demonstrated by several groups (Curry and Lohmann, 1982; Shackleton and others, 1983; Boyle and Keigwin, in press); for the late Miocene, by Bender and Graham (1981); and for the Oligocene to middle Miocene, by Miller and Fairbanks (1983 and 1985). Our paper lacks the long time series of *Cibicidoides* data necessary to extend these studies beyond the Oligocene. Rather, Figure 10a presents average late Eocene and early Oligocene benthic foraminiferal $\delta^{13}C$, normalized to *Cibicidoides*, as a function of paleodepth. There is considerable scatter in the data and no apparent difference between either time interval. Limited evidence exists for minimum $\delta^{13}C$ values at ~1,500 m in Atlantic Ocean locations. Such a pattern may reflect an Eocene/Oligocene oxygen minimum (and hence a minimum in $\delta^{13}C$ of ΣCO_2), which resulted from a warmer, well-mixed, less dense surface ocean covering a denser, probably cooler, deep ocean, and coupled with oxidation of organic matter as it sank from the surface layer. Some variability in benthic $\delta^{13}C$ data at a given water depth probably results from real differences in sea-water chemistry between deep
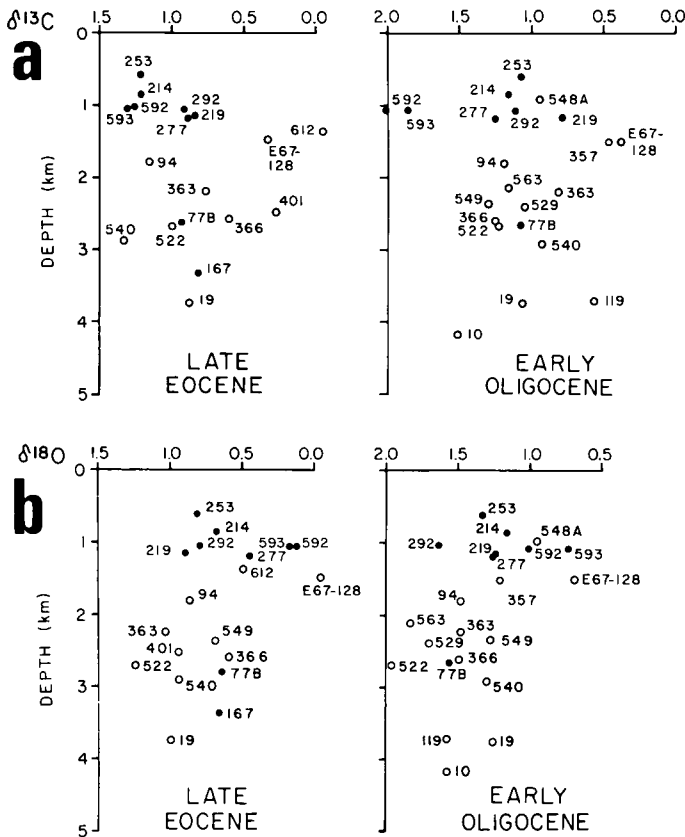
Figure 10. Average late Eocene and early Oligocene carbon (a) and oxygen (b) isotope data on benthic foraminifera plotted against the paleodepth 37 m.y. ago. Solid points are Indo-Pacific, and open points are Atlantic Ocean data. In the Atlantic samples, minimum $\delta^{13}C$ values at about 1,500 m may reflect the position of the oxygen minimum zone, and generally increasing $\delta^{18}O$ values with depth probably reflect decreased temperature.

oceans and between deep basins within an ocean. Other sources of scatter may be the result of errors in backtracking and errors in "correcting" of data to *Cibicidoides* values. None of these factors, however, could account for the extraordinary enrichment in ^{13}C of *Cibicidoides* from Sites 592 and 593. A final source of scatter is our averaging of data close to the time scale on which it is varying. In this regard, Miller and Fairbanks (1985) report that the lower Oligocene is marked by one-half a global cycle in $\delta^{13}C$, with an amplitude of $>1^o/oo$. We may not have always averaged data from the same part of that cycle, because the interval for averaging was chosen on the basis of maximum $\delta^{18}O$ values.

Oxygen-isotope results in benthic foraminifera, normalized to *Cibicidoides*, are presented in Figure 10b. It is evident that the main difference between Eocene and Oligocene results is the earliest Oligocene average enrichment of ^{18}O by $0.64^o/oo$ (from data in Table 2). Although the data are widely scattered, they reveal a general trend of increasing $\delta^{18}O$ with increasing paleodepth for each time interval. This trend is best developed among *Cibicidoides* data from Atlantic Ocean locations, where there are the most data and where $\delta^{18}O$ increases $\sim1^o/oo$ between 1,000 and 2,500 m. A similar $\delta^{18}O$ increase has been used previously to argue that the thermal structure of the Oligocene ocean (Douglas and Savin, 1978) and the early Pliocene ocean (Keigwin and others, 1979) was similar to that of today. Such patterns of $\delta^{18}O$ increase are to be expected with a density-stratified ocean, but we are unable to determine from this data alone if the increased density with depth results from decreased temperature or from increased salinity (see, for example, Brass and others, 1982). Independent evidence, however, suggests initial or increased production of cool northern-source bottom waters in the Atlantic Ocean (Miller and Tucholke, 1983) and southern-source bottom waters in the Pacific Ocean (Kennett and others, 1972) during Oligocene time.

Latitudinal Reconstructions

Previous studies have indicated the usefulness of latitudinal $\delta^{18}O$ gradients in planktonic foraminifera in describing the planetary temperature gradient (Savin and others, 1975; Shackleton and Boersma, 1981). Latitudinal variations in carbon isotopes in fossil planktonic foraminifera
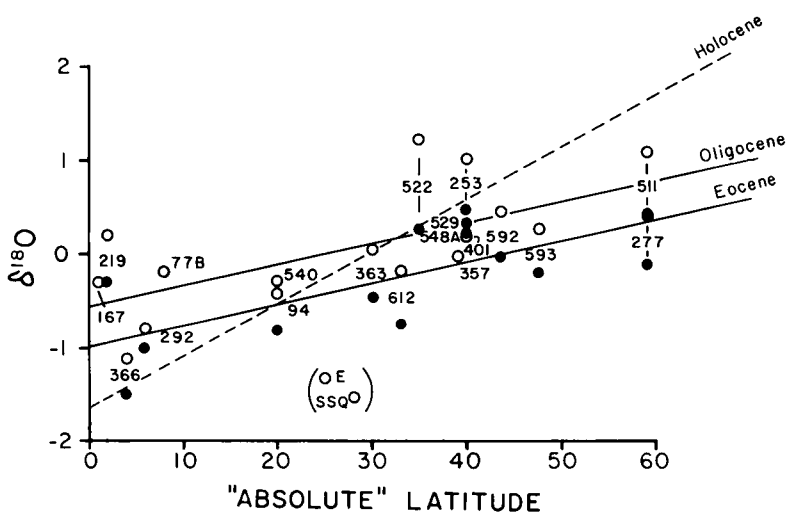

Figure 11. Average late Eocene and early Oligocene oxygen-isotope results in surface-dwelling planktonic foraminifera plotted against paleolatitude without respect for hemisphere ("absolute" latitude). Eocene (solid data points) and Oligocene (open data points) are fitted with regression lines with slopes of 0.023 (r = 0.71) and $0.022^o/oo$ (r = 0.69) per degree latitude, respectively. Data from E67-128 and from St. Stephens Quarry are not included in the regression analysis because of the low $\delta^{18}O$ for their latitude, probably reflecting elevated temperatures or lowered salinity and because E67-128 results may be affected by reworking (Poore and Matthews, 1984b). The Paleogene $\delta^{18}O$ latitudinal gradient is considerably less than the Holocene gradient ($0.056^o/oo$ per degree latitude) calculated from the data of Durazzi (1981). The significant difference between the Eocene and Oligocene regression lines is the increase in the y-intercept of $0.45^o/oo$, which probably reflects the effects of increased continental glaciation in earliest Oligocene time.

have not been successfully modeled, and results of this study seem too scattered to interpret (Table 2). Perhaps this results from a strong taxonomic influence on the $\delta^{13}C$ in planktonic foraminifera, as observed in data from North Atlantic core tops (Durazzi, 1981).

Most of our sites backtrack to southern latitudes, and there are too few sites to attempt determining gradients in stable isotope composition for each hemisphere (Table 2). For this reason, $\delta^{18}O$ data are plotted against latitude without regard for hemisphere ("absolute" latitude), assuming a bipolar temperature distribution, and fitted with regression lines (Fig. 11). Data from E67-128 and from St. Stephens Quarry, which probably reflect a low salinity influence or warmer waters associated with marginal seas, were excluded from the linear regression. Scatter about the regression lines in Figure 11 probably results from the few analyses per site, as well as from the effects of local temperature, salinity, and, possibly, diagenesis. On average, the $\delta^{18}O$ gradients represent sea-water density gradients, which are ultimately a function of the planetary temperature gradient. On the basis of this evidence, the planetary temperature gradient did not change appreciably across the Eocene/Oligocene boundary.

The most significant difference between the Eocene and Oligocene data in Figure 11 is the increase of almost $0.45\%_0$ in the y-intercept, which probably results from an increase in Oligocene continental ice volume. Significant continental ice accumulation as long ago as earliest Oligocene time has been argued on theoretical grounds (Matthews and Poore, 1980); it has also been previously acknowledged based on covariance patterns among oxygen-isotope ratios in benthic and planktonic foraminifera (Keigwin, 1980), and on the magnitude of the earliest Oligocene ^{18}O enrichment (Miller and Fairbanks, 1983; Poore and Matthews, 1984a; Shackleton and others, 1984). The covariance argument was used to show that most of the ^{18}O enrichment was due to deep-sea cooling, as indicated by large increases in Oridorsalis $\delta^{18}O$ coupled with smaller increases in planktonic foraminiferal $\delta^{18}O$ (Keigwin, 1980). Our present results, however, suggest that Oridorsalis may amplify the deep-sea $\delta^{18}O$ signal in the Oligocene, which would overemphasize the importance of cooling in the earliest Oligocene. It appears that, on average, deep-sea cooling may have been no more than 2 °C. One line of evidence is the small difference between the average benthic foraminiferal ^{18}O enrichment $(0.64\%_0)$ and the difference in y-intercepts in Figure 11 $(0.45\%_0)$. An upper limit of 2 °C cooling at some locations is the difference between the inferred sea-water compositional effect $(0.45\%_0)$ and the maximum benthic foraminiferal change $(0.9\%_0$ at Sites 77B, 366, and 592; Table 2). Although we now believe that the effects of Eocene/Oligocene glaciation in the $\delta^{18}O$ record are relatively more important than paleotemperature change, there is still ample paleontological evidence for cooler Oligocene climate (Haq and others, 1977; Keller, 1982/1983; Wolfe, 1978). As some locations in the deep sea may have cooled by 2 °C, high-latitude surface waters in the source region must also have cooled by at least this amount. Unfortunately, the present lack of suitable samples and sites from higher latitudes prevents quantifying a change in the planetary temperature gradient using the oxygen isotope method. The scarcity of sites poleward of 40° also makes it difficult to test the hypothesis that surface water in high latitudes had significantly lowered salinity (Poore and Matthews, 1984b).

Our evidence for insignificant change in the latitudinal $\delta^{18}O$ gradient at the Eocene/Oligocene boundary contrasts with the recent results of Murphy and Kennett (in press) from the Tasman Sea. These workers concluded that by early Oligocene time surface waters overlying Site 277 had cooled by 1.5 °C compared to those overlying Site 593. Murphy and Kennett, however, disregarded any ice-volume effects at this time and calculated surface-water temperatures based on a species (*G. angiporoides*), which was probably not a surface dweller. Our results from these sites show that, during the latest Eocene and earliest Oligocene, *Chiloguembelina* spp. from Site 277 were consistently more enriched in ^{18}O than those from Site 593 by $\sim 0.1\%_0$ (Table 2, Fig. 11).

Comparison of Paleogene $\delta^{18}O$ gradients to the Holocene gradient (Durazzi, 1981) shows that the Holocene $\delta^{18}O$-latitude gradient is more than twice the Eocene or Oligocene gradient (Fig. 11). The Holocene gradient was calculated from Durazzi's *Orbulina universa* data and is probably a minimum gradient. Although *O. universa* did not always have the lowest $\delta^{18}O$ value in each sample, data from this species spanned nearly 60° of latitude (similar to that of our data); Durazzi concluded that it is a species for which temperature exerts a strong control on $\delta^{18}O$. From a comparison of these results, we conclude that the Eocene and Oligocene latitudinal temperature gradients were half the modern gradient, at least within 60° of the equator.

CONCLUSIONS

Time series of middle Eocene to early Oligocene stable isotope data on foraminifera from Atlantic, Pacific, and Indian Ocean locations reveal stepwise increases in $\delta^{18}O$ in benthic foraminifera associated with the middle/late Eocene boundary and the Eocene/Oligocene boundary. A maximum in $\delta^{13}C$ occurs in the earliest Oligocene. At many locations, including those in the tropics, $\delta^{18}O$ of surface-dwelling planktonic foraminifera increased in the earliest Oligocene. This indicates some ice-volume increase at that time. At the middle/late Eocene boundary, $\delta^{18}O$ of surface-dwelling planktonic foraminifera does not increase, an indication of greater influence of bottom-water cooling. Glaciation cannot be ruled out, however, on the basis of observations at one location. There appears to be a change in the "vital effect" of the benthic foraminifera *Oridorsalis tener* which results in increased $\delta^{18}O$ compared to *Cibicidoides* spp. in Oligocene samples. This effect may have amplified the apparent bottom-water cooling inferred in earlier studies using *O. tener*. The $\Delta\delta^{18}O$ between these species as well as between pairs of *Cibicidoides* and *Uvigerina* is less in Paleogene time than in Neogene time, which means that Neogene and Quaternary "correction factors" cannot be used on Paleogene species.

Averaging stable isotope results is useful to emphasize the contrast in oceanographic and climatic conditions between late Eocene and early Oligocene time. The following results were revealed.

1. A change in the planetary temperature gradient from Eocene to Oligocene time is not evident from our comparison of planktonic foraminiferal $\delta^{18}O$ data versus paleolatitude.

2. Oxygen isotope data from planktonic foraminifera indicate an increase in sea-water $\delta^{18}O$ of $\sim 0.45\%_0$, probably due to growth of continental ice in early Oligocene time.

3. The average Eocene/Oligocene boundary change in benthic foraminiferal $\delta^{18}O$ is $\sim 0.6\%_0$. Using a sea-water compositional effect of $\sim 0.45\%_0$, the deep sea cooled in early Oligocene time by as much as 2 °C at some locations and by 1 °C overall.

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