PALEOCEANOGRAPHY, VOL. 2, NO. 1, PAGES 1-19, FEBRUARY 1987

TERTIARY OXYGEN ISOTOPE SYNTHESIS, SEA LEVEL HISTORY, AND CONTINENTAL MARGIN EROSION

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Abstract. Tertiary benthic and planktonic foraminiferal oxygen isotope records are correlated to a standard geomagnetic polarity time scale, making use of improved chronostratigraphic control and additional Oligocene isotope data. Synchronous changes in both benthic and planktonic $\delta^{18}O$ values which occurred in the Oligocene to Miocene (36-5.2 Ma) are interpreted, in part, to represent ice growth and decay. The inferred ice growth events correlate with erosion on passive continental margins as interpreted from seismic and chronostratigraphic records. This association is consistent with a link between Oligocene to Miocene erosional events and rapid (>15 m/m.y.) glacioeustatic lowerings of about 50 m. High benthic foraminiferal $\delta^{18}O$ values suggest the presence of continental ice sheets during much of the Oligocene to Recent (36-0 Ma). Substantially ice-free conditions probably existed throughout the Paleocene and Eocene (66-36 Ma). The mechanisms and rates of sea level change apparently were different between the early and late Tertiary, with glacio-eustatic changes restricted to the past 36 m.y. Pre-Oligocene erosion on passive continental margins was caused by eustatic lowerings resulting from global spreading rate changes. We apply a model which suggests that large areas of the continental shelves were subaerially exposed during such tectonoeustatic lowstands, stimulating slope failure and submarine erosion. The different mechanisms and rates of eustatic change may have caused contrasting erosional patterns between the early and late Tertiary on passive continental margins. This speculation needs to be confirmed by examination of data from several passive margins.

Paper number 6P0712.
0883-8305/87/006P-0712$10.00

INTRODUCTION

Ice volume changes dominated Pleistocene $\delta^{18}O$ fluctuations measured in deep-sea sediments [Shackleton and Opdyke, 1973]. Separating Tertiary ice volume from temperature effects on the oxygen isotope record has proven more difficult. The first detailed Tertiary $\delta^{18}O$ studies from deep-sea sediments assumed that the earth was substantially ice free prior to the middle Miocene (about 15 Ma) [Shackleton and Kennett, 1975a; Savin et al., 1975]. Later isotope studies argued for large continental ice sheets since at least the earliest Oligocene (Matthews and Poore [1980], Miller and Fairbanks [1983, 1985], and Keigwin and Keller [1984], among others). Evidence of upper Oligocene glaciomarine sediments in the Ross Sea confirms the presence of glacial ice on the margin of the Antarctic continent [Barrett et al., 1987], and many now accept evidence for Oligocene ice sheets [e.g., Shackleton et al., 1984a; Savin and Barrera, 1985]. Nevertheless, the initial onset, timing, and magnitude of Tertiary ice volume changes are poorly known.

Initial Tertiary stable isotope studies of piston cores and outcrop sections in the 1950s and 1960s noted that foraminiferal oxygen isotope values increased over the last 50 m.y., indicating that the oceans have cooled and continental ice sheets have expanded [Emiliani, 1954; Dorman, 1966; Devereux, 1967]. In the 1970s, recovery of deep-water sections by the Deep Sea Drilling Project (DSDP) enabled determination of detailed changes superimposed upon this general increase in $^{18}O/^{16}O$ ratios during the Tertiary. Shackleton and Kennett [1975a] and Savin et al. [1975] showed that following a $\delta^{18}O$ minimum in the early Eocene, sharp oxygen isotope increases occurred near the Eocene/Oligocene boundary and in the early middle Miocene. Subsequent studies improved the stratigraphic control on these $\delta^{18}O$ changes. For example, benthic foraminiferal $\delta^{18}O$ values increased at the beginning of the Oligocene, at approximately 36-35 Ma within Chron C13n, following extinction of foraminiferal species used to recognize the Eocene/Oligocene boundary (Hantkenina spp., Globorotalia cerroazulensis ssp., see Kennett and Shackleton [1976], Corliss et al. [1984], Oberhansli et al. [1984], and Miller et al. [1985a], among others). Also, the middle Miocene $\delta^{18}O$ increase occurred in benthic foraminifera from the Atlantic, Pacific, and Indian Oceans in the early middle Miocene associated with Zones N9/N10 and NN5/NN6 at approximately 15-13 Ma (see Savin et al. [1981], Woodruff et al. [1981], Miller and Fairbanks [1983, 1985], and Vincent et al. [1985], among others).

Previous Tertiary benthic foraminiferal $\delta^{18}O$ syntheses were limited by lack of complete Oligocene sections, problems in correlations based solely upon biostratigraphy, and uncertainties in time scales used in the syntheses. We believe that a new synthesis is justified which includes data from continuous Oligocene strata [Miller and Fairbanks, 1983, 1985; Keigwin and Keller, 1984; Miller and Thomas, 1985; this study], takes advantage of improved stratigraphic control afforded by magnetostratigraphy [e.g., Tauxe et al., 1984; Miller et al., 1985b], and applies a single improved magnetobiostratigraphic time scale for all age correlations [Berggren et al., 1985]. We have assembled a composite record using only the most complete individual benthic foraminiferal $\delta^{18}O$ time series obtained on the most reliable taxa [see "methods" section] from DSDP sections in order to document stable isotope fluctuations on the 10^6- to 10^7-year scale. We present individual compilations for the Atlantic (Figure 1) and Pacific (Figure 2). Our compilations differ from earlier compilations [e.g., Savin, 1977; Shackleton et al., 1984a] because our resolution is improved over certain critical intervals. For example, we resolve two Oligocene oxygen isotope fluctuations that were not recongized because of hiatuses and coarser sampling in previous work.

Tertiary ice volume history is estimated by using benthic and low- to middle-latitude planktonic foraminiferal $\delta^{18}O$ records. This glacioeustatic history is compared with chronostratigraphic and seismic stratigraphic records of erosion on passive continental margins. We suggest that the mechanisms for eustatic changes and continental margin response to these changes differed between the early and late Tertiary.

METHODS

Foraminiferal $\delta^{18}O$ changes reflect temperature and global seawater compositional (ice volume) changes as expressed in the paleotemperature equation:

$$T = 16.9 - 4.38(\delta_c - \delta_w) + 0.10(\delta_c - \delta_w)^2$$

where T is paleotemperature, δ_w is global seawater composition, and δ_c is the measured value in calcite assuming equilibrium calcification [O'Neil et al., 1969; Shackleton, 1974]. We have attempted to differentiate Tertiary ice volume changes from temperature fluctuations by making the following assumptions:

1. Benthic foraminiferal values can be used to indicate when the world was glaciated by assuming that cold (less than about 2°C) bottom waters indicate that high latitudes were frigid enough to support large ice sheets [e.g., Miller and Fairbanks, 1983; Keigwin and Keller,

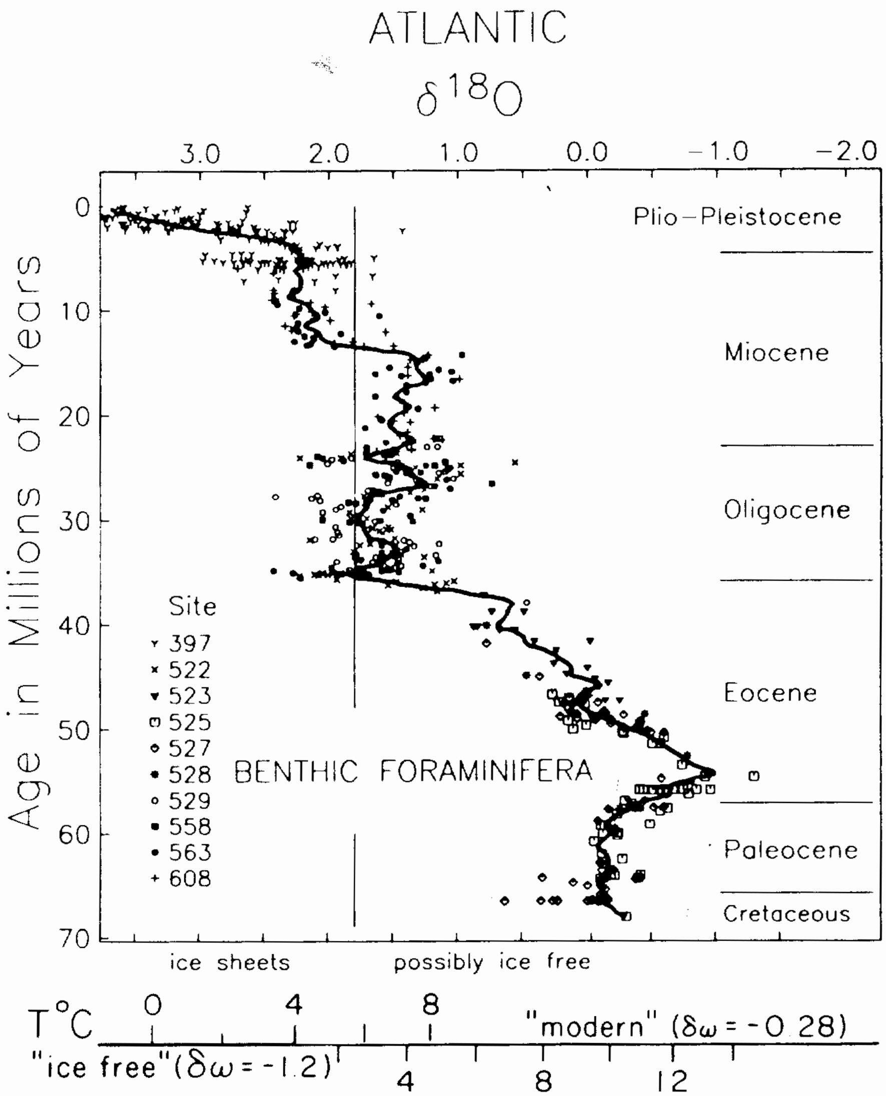

Fig. 1. Composite benthic foraminiferal oxygen isotope record for Atlantic DSDP sites (Table 1) corrected to Cibicidoides (see "methods" section) and reported to Pee Dee Belemnite standard (PDB). Chronostratigraphic subdivisions are drawn after Berggren et al. [1985]. The smoothed curve is obtained by linearly interpolating between data at 0.1-m.y. intervals and smoothing with a 27-point Gaussian convolution filter, removing frequencies higher than 1.35/m.y. The vertical line is drawn through 1.8°/oo; values greater than this provide evidence for existence of significant ice sheets. The temperature scale is computed using the paleotemperature equation, assuming Cibicidoides are depleted relative to equilibrium by 0.64°/oo. The lower temperature scale assumes no significant ice sheets, and therefore δ_w = -1.2°/oo; the upper scale assumes ice volume equivalent to modern values, and therefore δ_w = -0.28°/oo.

1984]. Benthic foraminiferal $\delta^{18}O$ records therefore can be used to indicate the presence of ice sheets but not the timing or magnitude of ice volume changes.

2. Ice growth and decay causes global seawater compositional (δ_w) changes that produce synchronous $\delta^{18}O$ changes in both benthic and planktonic foraminifera. Such synchronous $\delta^{18}O$ changes in benthic and low- to middle-latitude planktonic foraminifera may potentially indicate ice growth and decay events; applications of this method are provided by Shackleton and Opdyke [1973] and Crowley and Matthews [1983] for the Pleistocene, by Prell [1984] for the Pliocene, and by Matthews

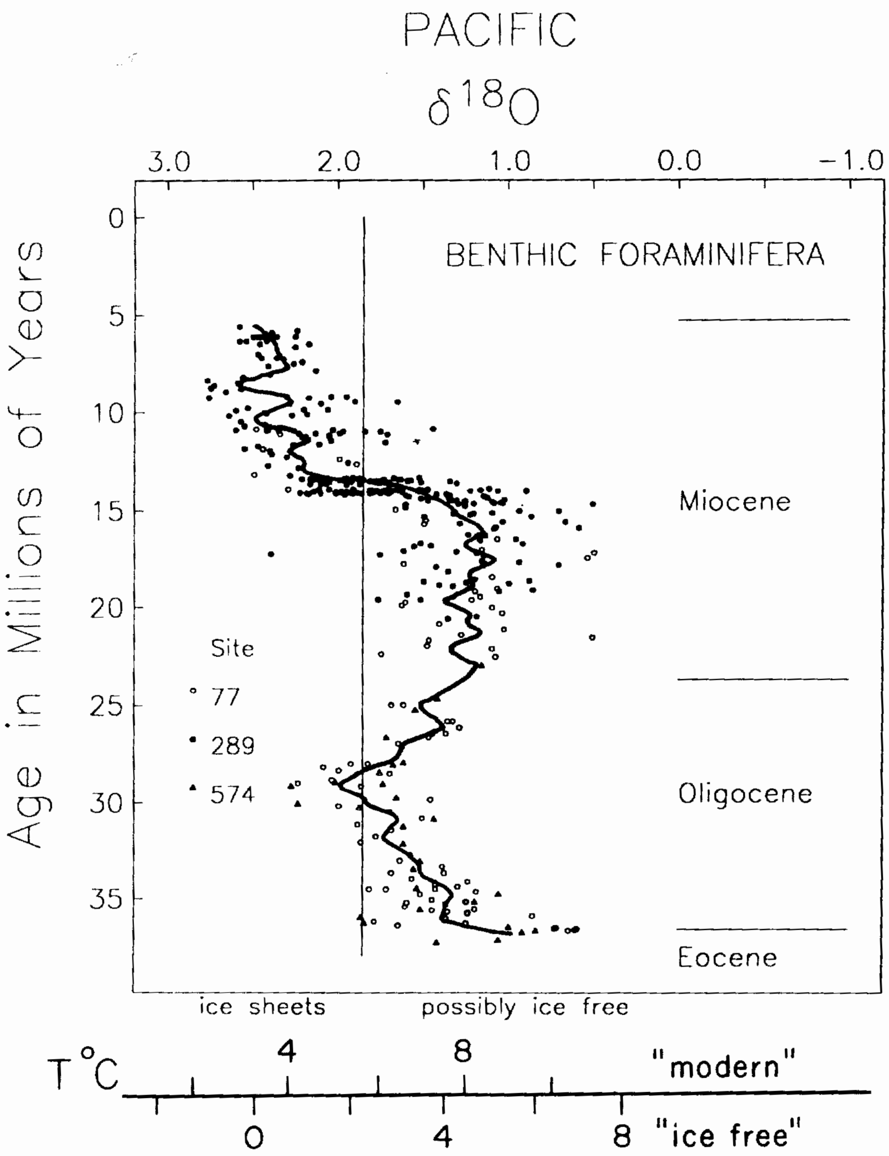

Fig. 2. Composite benthic foraminiferal (all Cibicidoides) oxygen isotope record for Pacific DSDP sites (Table 1) reported to PDB standard. Chronostratigraphic subdivisions and temperature scale are drawn as in Figure 1. The smoothed curve is obtained by linearly interpolating between data at 0.1-m.y. intervals and smoothing with a 21-point Gaussian convolution filter, removing frequencies higher than 1.05/m.y.

and Poore [1980] and Miller and Fairbanks [1985] for the Oligocene. However, such synchronous changes may also be interpreted as concomitant bottom water and low- to middle-latitude surface water temperature changes [e.g., Savin, 1977]. We assume that ice growth is indicated by increased benthic and planktonic $\delta^{18}O$ values followed by low bottom water temperatures (<2°C). We have restricted comparisons of benthic and planktonic foraminiferal oxygen isotope records to the Oligocene and Miocene because Paleocene to Eocene planktonic foraminiferal records are relatively poor.

Age Control

Age control is provided by magnetostratigraphy and biostratigraphy. Previously published Tertiary oxygen isotope records were dated using several different time scales. We standardized all isotope data to the geomagnetic polarity time scale of Berggren et al. [1985]. Age models for all Atlantic sites were derived from magnetobiostratigraphy (e.g., Sites 522, 523, 525, 527, 528, 529, 558, 563, and 608 [Hamilton, 1979; Tauxe et al., 1984; Shackleton et al., 1984b; Miller et al., 1985b; Clement and Robinson, 1987; K.G. Miller and D.V. Kent, unpublished data, 1986]). Magnetobiostratigraphic age models for Sites 523-528 follow Shackleton et al. [1984b], who also used the Berggren et al. [1985] time scale. Indo-Pacific sites were correlated to the time scale using biostratigraphy (e.g., Sites 77, 214, 216, 237, 253, 289, and 574). The isotope data and age model parameters are available upon request.

Benthic Record

We compiled benthic foraminiferal Tertiary $\delta^{18}O$ records from the Atlantic (Figure 1; Table 1) and Pacific Oceans (Figure 2; Table 1), using data from sites with paleodepths greater than 2 km (except as noted in Table 1). The Atlantic contains a particularly good Paleogene record (Figure 1). The Pacific contains a good Oligocene to Miocene record (Figure 2), while Paleocene to Eocene coverage is poor and is not included here. Atlantic data from 35 to 8 Ma and all Pacific data are based upon analyses of Cibicidoides spp.; this taxon secretes calcite which is apparently offset from $\delta^{18}O$ equilibrium by about 0.64°/oo (Shackleton and Opdyke [1973], Graham et al. [1981], among others). For the late Miocene to Recent (8-0 Ma), we have included analyses of Uvigerina spp.; this taxon precipitates its test near oxygen isotope equilibrium with ambient seawater [Graham et al., 1981]. The Paleocene to Eocene (65-36 Ma) Atlantic record is based upon analyses of monogeneric samples (see Shackleton et al. [1984a] for taxa from Sites 525-527 and Oberhansli et al. [1985] for taxa from Site 523); the departures of these taxa from isotope equilibrium are less well known. We adjusted all data to Cibicidoides using data of Shackleton et al. [1984a] except as noted in Table 1.

The oxygen isotope data obtained (Figures 1 and 2) show a high degree of scatter (typically 0.5-1.0°/oo) that we attribute to uncertainties in stratigraphic correlations and combining data from different hydrographic regimes. In addition, some of the scatter may represent undersampling of high-frequency (10^4 to 10^5 year) signals. High-frequency "Milankovitch" ice volume signals have been documented for the latest Cenozoic [Hays et al., 1976; Imbrie et al., 1984], and high-frequency (10^5 year) oxygen isotope changes have been observed in the middle Miocene [Shackleton, 1982; Pisias et al., 1985] and possibly the earliest Oligocene [Poore and Matthews, 1984]. We smoothed the time series by interpolating the data to a constant 10^5-year interval (equivalent to the average Oligocene to Miocene sampling intervals). Frequencies higher than about 1/m.y. were eliminated using Gaussian convolution filters (27-point filter for the Atlantic; 21-point filter for the Pacific). The filters are Gaussian-shaped running means of about 1 m.y. (1.05 m.y. for the Pacific; 1.35 m.y. for the Atlantic; greater smoothing was required for the Atlantic because of the greater number of sites combined); small changes in the smoothed curves (<0.3°/oo) are probably not signficant. Larger changes in the smoothed curves delineate isotope changes that have been recognized previously (e.g., the earliest Oligocene increase) and new features that are less well known (middle and latest Oligocene peaks).

Planktonic Record

We synthesized tropical and subtropical planktonic foraminiferal $\delta^{18}O$ records from the Atlantic and Indian Oceans. Atlantic records are based upon DSDP Sites 522 (26°S present latitude), 563 (33°N), and 558 (38°N), which are located in subtropical gyres. Isotope analyses at Sites 563 and 558 were performed on Globorotalia opima nana [Miller and Fairbanks, 1985]; analyses at Site 522 were performed on Globigerina euaperta and Globigerina pseudovenezuelana [Oberhansli et al., 1984; Poore and Matthews, 1984]. These taxa are among the most depleted in ^{18}O at these locations (i.e., within 0.3°/oo of most depleted taxa measured), and we assume that they calcified in the surface mixed layer [Poore and Matthews, 1984; Miller and Fairbanks, 1985]. Indian Ocean Miocene records from Sites 214 (11°S present latitude, 20°S paleolatitude), 216 (1°N present latitude, 7°S paleolatitude), and 237 (7°S present latitude, 9°S paleolatitude) are based upon anlyses of

TABLE 1. Sources for Benthic Foraminiferal Oxygen Isotope Data

Site	Location	Taxon	Age	Reference
		Atlantic Sites		
397	off west Africa 26°50.7'N, 15°10.8'W 2900 m	Cibicidoides, Uvigerina	late Miocene to Recent	Shackleton and Cita [1979], Stein [1984]
522	eastern South Atlantic 26°6.8'S, 05°7.8'W 4441 m	Cibicidoides	latest Eocene to Oligocene	Poore and Matthews [1984], this study
523	eastern South Atlantic 28°33.1'S, 02°15.1'W 4562 m	Nuttallides, Oridorsalis	middle to late Eocene	Oberhansli et al. [1984]
525	Walvis Ridge 29°04.2'S, 02°59.1'E 2467 m; 1600 m paleo depth at 47 Ma, 900m paleodepth at 64 Ma	various	Cretaceous to middle Eocene	Shackleton et al. [1984a]
527	Walvis Ridge 28°02.5'S, 01°45.8'E 4428 m	various	Cretaceous to middle Eocene	Shackleton et al. [1984a]
528	Walvis Ridge 28°31.5'S, 02°19.4'E 3815 m	various	Eocene to earliest Oligocene	Shackleton et al. [1984a]
529	Walvis Ridge 28°55.8'S, 02°46.1'E 3055 m	Cibicidoides various	Oligocene; Eocene to early Oligocene	this study; Shackleton et. al. [1984]
558	western North Atlantic 37°46.2'N, 37°20.6'W 3754 m	Cibicidoides	Oligocene to Miocene	Miller and Fairbanks [1985]
563	western North Atlantic 33°38.5'N, 43°46.0'W 3793 m	Cibicidoides	Oligocene to Miocene	Miller and Fairbanks [1985]
608	eastern North Atlantic 42°50.2'N, 23°05.3'W 3526 m	Cibicidoides	Miocene	Miller et al. [1987]
		Pacific Sites		
77	equatorial Pacific 00°28.9'N, 133°13.7'W 4291 m	Cibicidoides	Miocene; Oligocene	Savin et al. [1981]; Keigwin and Keller [1984]
289	equatorial Pacific 00°29.9'S, 158°30.7'E 2206 m	Cibicidoides	Miocene	Savin et al. [1981]
574	equatorial Pacific 04°12.5'N, 133°19.8'W 4536 m	Cibicidoides	Oligocene	Miller and Thomas [1985]

All values are adjusted to Cibicidoides. Correction factors and age models for Sites 525, 527, and 528 are after Shackleton et al. [1984a, b]; eastern North Atlantic Site 608 Miocene data differ from Miocene records from the western North Atlantic because of temperature and/or salinity differences [Miller et al., 1987]. Site 608 data were calibrated to western Atlantic data by adding 0.42°/oo, the difference between the mean of western basin and the eastern basin for the Miocene.

Globigerinoides sacculifer and Globigerinoides altispira [Vincent et al., 1985]. The Eocene to Oligocene Indian Ocean record is based upon subtropical Site 253 (25°S present latitude, 36°S paleolatitude)[Keigwin and Corliss, 1986; Oberhansli, 1986] on the following taxa: Globigerina eocaena, Globigerina pseudovenezuelana, Chiloguembelina cubensis, and Pseudohastigerina micra, which are all assumed to dwell in the surface mixed layer [Poore and Matthews, 1985]. Paleolatitudes for the Indian Ocean sites are taken from Sclater et al. [1985](Sites 214, 216, 237 from the 16 Ma reconstruction; Site 253 from 22 Ma reconstruction). Paleolatitudues of Atlantic sites are similar to the present latitudes [Sclater et al., 1985].

Diagenesis

Diagenesis may overprint or erase the isotope composition originally recorded by foraminifera and create artifacts in the record. For example, isotopic records at deeply buried (>400 m) sites may be depleted by up to 3.0°/oo relative to data from unaltered samples [Miller and Curry, 1982], suggesting complete removal of the original signal. Such obviously anomalous isotope records were avoided here. However, diagenetic effects may be more subtle and pervasive. Killingley [1983] ascribed much of the first-order $\delta^{18}O$ increase of the Cenozoic (Figures 1 and 2) to increased diagenesis with burial depth. Since major features of the oxygen isotope record agree among cores with different burial depths and thermal histories, Killingley argued that the original isotope signal may not have been completely erased, only overprinted.

We tested whether increased oxygen isotope values during the last 50 million years are partially due to increased diagenesis with burial depth and found that diagenetic overprinting is generally minor in the sections selected for this paper. We compiled Miocene benthic foraminiferal isotope records from different burial depths ranging from 100 to 600 m into two Miocene time intervals: 5-15 Ma and 15-25 Ma (Figure 3), and compared them with the relationship predicted by the Killingley [1983] model. The oxygen isotope data show no systematic relationship with burial depths values for these Miocene intervals contrary to predictions of the Killingley model (Figure 3). Similarly, for Paleocene to lower middle Eocene sections (closed circles, Figure 3, right column) there is no relationship between burial depths less than 500 m, isotope composition, and the Killingley model. Deeper than 500 m, some Eocene values are quite depleted in ^{18}O; these data come from sections already suspected to be severely altered [Miller and Curry, 1982].

OXYGEN ISOTOPE CHANGES ON THE 10^6- TO 10^7-YEAR SCALE

The benthic foraminiferal Atlantic synthesis shows that oxygen isotope values generally increased from the Eocene to Recent (Figure 1), in agreement with previous work. Oxygen isotope values decreased across the Paleocene/Eocene boundary to a distinct minimum in the early Eocene (approximately 57 to 52 Ma)(Figure 1). Early Eocene values (approximately -0.5°/oo) constrain bottom water temperatures to 11°C to 15°C. The lower temperature estimates given throughout were computed using the paleotemperature equation, mean $\delta^{18}O$ values for the given interval, and δ_w = -1.2°/oo; this δ_w value is that of an ice-free world [Shackleton and Kennett, 1975a]. The higher temperature estimates use δ_w = -0.28°/oo in the paleotemperature equation, thereby assuming ice volume equivalent to modern ice sheets [Craig, 1965; Craig and Gordon, 1965]. We assumed that Cibicidoides is depleted relative to isotope equilibrium by 0.64°/oo [Graham et al., 1981].

If bottom waters formed at high latitudes in the early Eocene as they do today, the inferred warm bottom water temperatures imply warm high latitude surface temperatures and ice-free conditions in and near polar coastal regions. Brass et al. [1982] suggested that warm, high-salinity bottom water might have formed at low latitudes in the past; if so, then benthic foraminiferal oxygen isotope data do not require warm poles [e.g., Matthews and Poore, 1980]. However, warm early Eocene polar conditions also are suggested by planktonic foraminiferal oxygen isotope evidence. At Site 277 (52°S), low early Eocene planktonic $\delta^{18}O$ values (mean of -1.2°/oo) may be interpreted as warm surface water temperatures (21°C, assuming δ_w = -0.28°/oo, to 17°C assuming δ_w = -1.2°/oo)[Shackleton and Kennett,

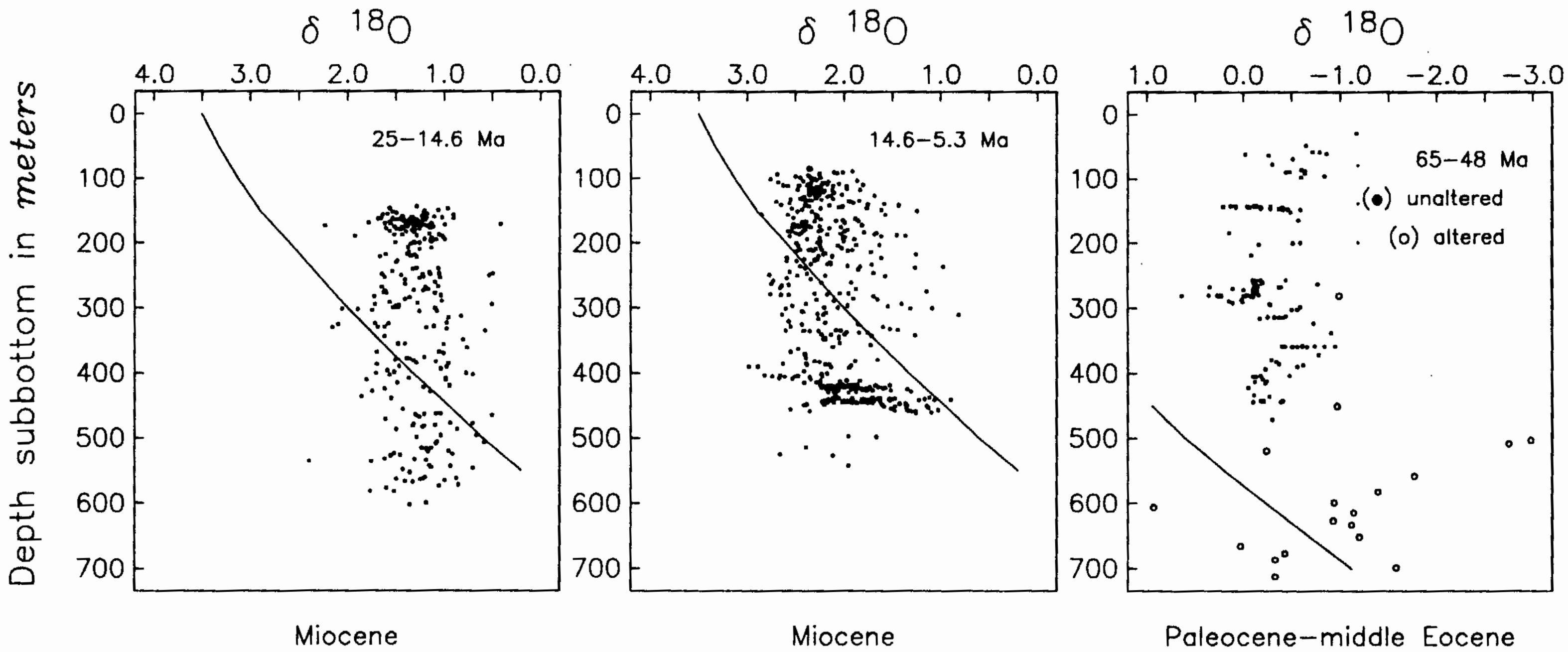

Fig. 3. Oxygen and carbon isotope values for DSDP sites (Table 1) divided into three time intervals: latest Oligocene to early middle Miocene (25-14.6 Ma), middle to late Miocene (14.6-5.3 Ma), and Paleocene to middle Eocene (65-48 Ma). The solid line is the predicted open system diagenetic relationship of Killingley [1983] (his model number 1). Open circles on the 65- to 48-Ma time interval represent data from Sites 112 [K.G. Miller and W.B. Curry, unpublished data, 1982), 398, 400 [Vergnaud-Grazzini et al., 1978], and 549 [Miller et al., 1985a] which are diagenetically altered. The lack of a trend with burial depth in meters subbottom argues against a simple depth-diagenetic relationship.

1975a]. Other paleoclimatic data also suggest warm high latitudes [Kemp, 1978; Wolfe, 1978; McKenna, 1980]. Therefore we believe that the assumption of a substantially ice-free world is valid for the early Eocene.

Oxygen isotope values increased between 52 and 40 Ma (middle Eocene)(Figure 1). This increase apparently occurred in two steps: (1) an increase of approximately 1.0^o/oo occurred between 52 and 48 Ma, beginning near the early/middle Eocene boundary; (2) an increase of approximately 0.5-0.7^o/oo occurred in the late middle Eocene (approximately 45-40 Ma), although the age control is poorly constrained (see Keigwin and Corliss [1986] for further discussion). Between 48 and 40 Ma (late middle Eocene), the average benthic $\delta^{18}O$ value (0.25^o/oo) indicates bottom water temperatures of 8^o to 12^{o}C. Typical values for the late Eocene (approximately 0.75^o/oo) indicate bottom water temperatures of 6^o to 10^{o}C. Benthic foraminiferal values of the middle to late Eocene (52-36 Ma) again suggest warm high-latitude surface temperatures. Terrestrial plant evidence also suggests warm and substantially ice-free conditions in high latitudes (e.g., Alaska [Wolfe and Hopkins, 1967]) during the middle and late Eocene. Thus we suggest that the $\delta^{18}O$ increases were due primarily to decreases in bottom water temperature.

Benthic foraminiferal isotope evidence indicates the presence of significant continental ice sheets in the Oligocene. Following the sharp $\delta^{18}O$ increase at 36-35 Ma (immediately after the Eocene/Oligocene boundary), values exceeded a 1.8^o/oo threshold (Figure 1), indicating that the earth was glaciated, presumably in Antarctica. To assume ice-free conditions when Cibicidoides $\delta^{18}O$ values are greater than 1.8°/oo would require bottom water and high-latitude surface waters to be colder than modern bottom water (Figure 1)(modern bottom water temperatures at Atlantic locations are typically 2.3-2.8^{o}C potential temperature [Fuglister, 1960]). For example, mean earliest Oligocene Atlantic $\delta^{18}O$ values (2.0^o/oo from 35.8 to 35.1 Ma) correspond to bottom water temperatures of 1.6^{o}C assuming an ice-free world, i.e. 1^{o}C colder than at present (using the paleotemperature equation, assuming that δ_w = -1.2^o/oo and that Cibicidoides are offset from equlibrium by 0.64°/oo). Such cold high-latitude temperatures are incompatible with an ice-free world. Since the modern $\delta^{18}O$ value at our Atlantic locations would be 1.8^o/oo if the present-day ice sheets were melted (i.e., assuming δ_w = -1.2^o/oo and modern Cibicidoides values of 2.7^o/oo), we use 1.8°/oo as the threshold for significant ice sheets (Figure 1). The threshold value actually lies between 1.6 and 1.9^o/oo depending upon the values selected for modern benthic foraminifera and δ_w.

Intervals of high benthic oxygen isotope values [>1.8^o/oo] occurred at approximately 31-28 Ma (Figures 1 and 2) and 25-24 Ma (Figure 1). The latter event is not resolved in the Pacific (Figure 2) owing to insufficient sampling. The event at approximately 31 to 28 Ma straddled the early/late Oligocene boundary, while the 25 to 24 Ma event occurred immediately before the Oligocene/Miocene boundary [Miller and Fairbanks, 1985]. Values decreased or remained stable from 24 to 16 Ma (early Miocene)(Figure 2). Benthic foraminiferal data do not unequivocally constrain whether ice sheets existed, and it is possible that large ice sheets were absent in the early Miocene [Shackleton and Kennett, 1975a; Savin et al., 1975, 1985; Woodruff et al., 1981]. The sharp $\delta^{18}O$ increase at approximately 15 to 13 Ma (Figures 1 and 2) undoubtedly reflects reestablishment or intensification of glacial conditions. High $\delta^{18}O$ values occurred at approximately 10 to 8 Ma (near the middle/late Miocene boundary)(Figures 1 and 2) and apparently at about 5.5-5 Ma (latest Miocene to earliest Pliocene)(for discussion of the former see Burckle et al. [1982]; for discussions of the latter see Elmstrom and Kennett [1985] and Keigwin et al. [1987]). The interval from 5.0 to 2.5 Ma (early Pliocene) is marked by lower oxygen isotope values [Elmstrom and Kennett, 1985]. This was followed by increased $\delta^{18}O$ variations beginning at 2.5-2.4 Ma in response to a major phase of northern hemisphere ice growth [Shackleton et al., 1984c].

Simultaneous oxygen isotope increases in bottom water and low- to middle-latitude surface waters can be interpreted as periods of ice growth (see the "methods" section for limitations of this approach). Synchronous changes between benthic and planktonic oxygen isotope records occurred several times during the Oligocene and Miocene (Figure 4):

1. Values increased in benthic and planktonic foraminifera immediately after

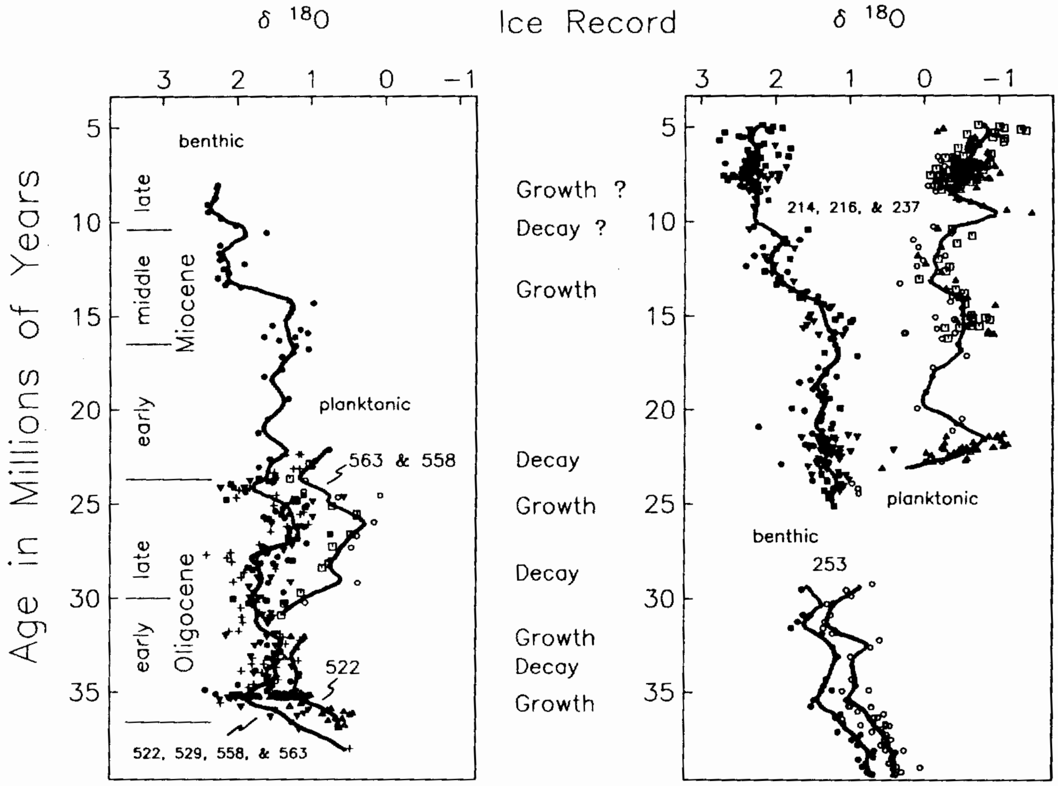

Fig. 4. Time series of benthic and surface-dwelling planktonic foraminifera. Ice growth (decay) events are indicated by synchronous increases (decreases) in both benthic (solid symbols) and planktonic (open symbols) foraminifera. Atlantic data from Sites 522 (triangles), 529 (pluses), 563 (circles), and 558 (squares) have been interpolated to a constant 0.2-m.y. sampling interval and smoothed with an 11-point Gaussian convolution filter, eliminating frequencies greater than 1/m.y. Indian Ocean Miocene data from Sites 216 (circles), 214 (squares), and 237 (triangles) [Vincent et al., 1985] have been interpolated to 0.1 m.y. sampling interval and smoothed with a 21-point filer, again eliminating frequencies greater than 1/m.y. Indian Ocean Eocene to Oligocene data (Site 253) [Keigwin and Corliss, 1986; Oberhansli, 1986] have been interpolated to 0.25-m.y. sampling interval and smoothed with a 9-point filter; the age model for this site is arbitrary. Indian Ocean Miocene benthic data generated on Oridorsalis [Vincent et al., 1985] have been corrected to Cibicidoides by subtracting 0.64°/oo; all other benthic data are based upon Cibicidoides.

the Eocene/Oligocene boundary at Sites 522 and 253 (approximately 36 Ma; Figure 4). At Site 292, (equatorial Pacific, Phillipine Sea), the increase in planktonic $\delta^{18}O$ values was less than that at Sites 522 and 253 (mean increase of about 0.2-0.3°/oo versus 0.5°/oo), although Site 292 was discontinuously cored, limiting resolution [Keigwin, 1980; Keigwin and Corliss, 1986].

2. In the "middle" Oligocene (about 31 Ma) benthic foraminiferal $\delta^{18}O$ values increased at many locations (Figures 1, 2, and 4). Planktonic records are sparse in this interval. Planktonic foraminiferal $\delta^{18}O$ values increased at temperate-subtropical Site 253 in the "middle" Oligocene (Figure 4), although age control is poor at this site and the site lay at fairly high latitude. Keigwin and Keller [1984] reported some increase in equatorial planktonic foraminiferal $\delta^{18}O$ records (Site 77), although details of this record are obscured by dissolution. A complete record from a low-latitude location is needed to confirm that the increase in planktonic foraminifera is not restricted to the middle latitudes. An

inferred ice volume decrease at about 28 Ma is interpreted from Atlantic and Indian Ocean sites (Figure 4).

3. Benthic and planktonic foraminiferal $\delta^{18}O$ values increased in the latest Oligocene (approximately 25 Ma) at North Atlantic sites (Figure 4). The planktonic increase began before the benthic increase, suggesting a surface water cooling in the subtropical North Atlantic prior to ice growth.

4. An increase in benthic foraminiferal values began in the middle Miocene at about 15 Ma. Planktonic and benthic values both increased from about 14.5 to 14.0 Ma. We interpret this as a bottom water temperature drop preceding ice growth (see also Kennett [1985] and Vincent et al. [1985]).

5. A period of possible ice growth occurred near the middle/late Miocene boundary (approximately 10-8 Ma). Western North Atlantic records are sparse at this time, whereas the Indian Ocean records either contain a hiatus or lack benthic data across this critical interval (Figure 4).

Comparison of planktonic and benthic oxygen isotope records (e.g., Figure 4) suggests covariance of at least 0.3-0.5°/oo at these times, which we interpret as estimates of changes in seawater $\delta^{18}O$. The relationship between seawater $\delta^{18}O$ and ice volume (hence sea level) is probably nonlinear [Mix and Ruddiman, 1984]. The Quaternary sea level- $\delta^{18}O$ calibration (0.11°/oo $\delta^{18}O$ per 10 m of sea level change) measured by Fairbanks and Matthews [1978] is an average value for relatively large ice volumes. Increasing area and altitude of growing ice sheets leads to lower mean ice sheet $\delta^{18}O$ values. Thus the early, smaller stages of ice sheet accumulation will have higher mean isotope values than those of later, larger ice sheets [e.g., Mix and Ruddiman, 1984; Savin and Douglas, 1985]. The maximum $\delta^{18}O$ values for snow accumulating around the modern Antarctic continent is approximately -17°/oo (SMOW) [Morgan, 1982]. Assuming this as a likely upper (most positive) limit for accumulation of ice sheets (versus about -35°/oo mean ice sheet composition today), then the sea level-$\delta^{18}O$ calibration would be 0.055°/oo per 10 m of sea level change. Therefore, small changes in seawater $\delta^{18}O$ reflect larger accumulations of ice during the early stages of ice growth than during later stages. For example, a 0.1°/oo change in foraminiferal $\delta^{18}O$ during the early stages of continental ice growth might have corresponded to a 20-m sea level lowering, while the same $\delta^{18}O$ change may equate to a 10-m sea level lowering during late stage growth of a large ice sheet.

The nonlinear oxygen isotope-sea level relationship may partially account for discrepancies between relatively small Oligocene to Miocene sea level changes estimated using isotope data and other estimates of Tertiary sea level lowerings. Using the Quaternary calibration and planktonic-benthic covariance estimate of 0.3-0.5°/oo, sea level was glacioeustatically lowered by 30-50 m at ca. 35, 31, 25, 14, and 10 Ma. Estimates made on coral atolls suggest a "middle" Oligocene lowering of about 100 m [Schlanger and Premoli Silva, 1986], while measurements of "offlap" [Vail et al., 1977; Vail and Hardenbol, 1979] suggest an even larger "middle" Oligocene lowering. Although the Quaternary sea level-$\delta^{18}O$ calibration suggests maximal lowerings of 50 m, sea level may have fallen as much as 90 m (assuming 0.5°/oo $\delta^{18}O$ change and calibration of 0.055). The 90-m estimate is a maximum upper limit, since ice would form at more negative values as the surface of the ice sheet grew to higher elevations. Thus, the glacioeustatic lowerings were more than 30 m and less than 90 m; they occurred in intervals of less than 2 m.y., yielding fairly rapid (>25 m/m.y.) rates of fall.

Although the benthic foraminiferal oxygen isotope data (Figure 1) allow an ice-free early Miocene interpretation, high-frequency (10^5 year) benthic-planktonic $\delta^{18}O$ covariance has been noted at this time [Shackleton, 1982], perhaps suggesting that rapid ice growth and decay may have occurred in the early Miocene.

EROSIONAL EVENTS AND THE $\delta^{18}O$ RECORD

We have evaluated chronostratigraphic and seismic stratigraphic evidence for Tertiary erosional events on continental margins and compared this evidence with the $\delta^{18}O$ record. Erosional events on passive continental margins are represented by unconformities, which can be identified using chronostratigraphy and seismic stratigraphy. Vail et al. [1977]

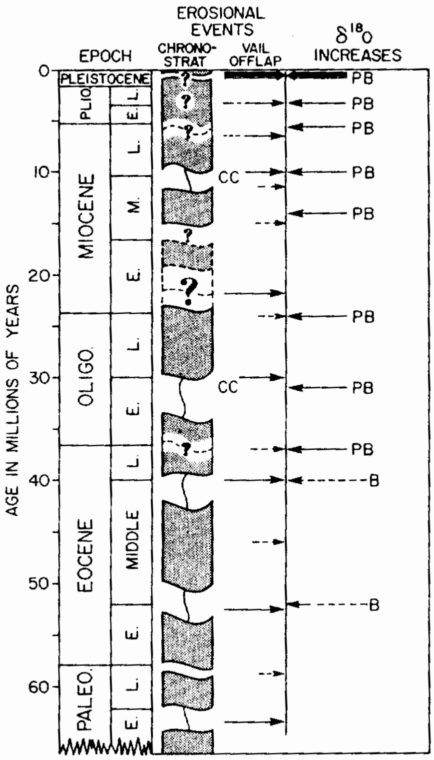

Fig. 5. Comparison of timing of Cenozoic erosional and isotopic events plotted using a uniform time scale [Berggren et al., 1985]. The chronostratigraphic column is a composite section from the U.S. Atlantic margin, the Irish margin (Goban Spur), and northwest European sections [Aubry, 1985; Melillo, 1985; Miller et al., 1985a, c; Poag et al., 1985; Snyder and Waters, 1985; Miller and Hart, 1987; Olsson and Wise, 1987]. Large arrows under "Vail offlap" events indicate the major, type 1 erosional events of Vail et al. [1977], Vail and Hardenbol [1979], and Vail and Mitchum [1980]; small arrows indicate smaller, type 2 events. PB denotes increases in both benthic and planktonic oxygen isotopes (Figure 4); B denotes an increase in benthic foraminifera. CC denotes canyon cutting events discussed by Miller et al. [1985c] and Farre [1985].

and Vail and Hardenbol [1979] suggested that seismic reflection profiles provide an accurate record of changes in sea level. However, the direct relationship between seismic stratigraphic record and global sea level changes has been challenged on at least two grounds: (1) maximum erosion apparently correlates with the greatest rate of sea level fall, not with lowest stand of sea level [Pitman, 1978; Watts, 1982; Miller et al., 1985c]; and (2) coastal onlap observed in seismic profiles can be of tectonic origin, unrelated to rising sea level (see discussion by Watts and Thorne [1984] and Christie-Blick et al. [1987]). Nevertheless, the major "offlap" (= downward shift in coastal onlap) events noted by Vail et al. [1977] in our estimation represent major erosional events on passive continental margins.

We compared the major ("Type 1") "offlap" events inferred from seismic stratigraphy by Exxon researchers [Vail et al., 1977; Vail and Hardenbol, 1979](Figure 5) with evidence for major chronostratigraphic breaks in continental slope and epicontinental settings (Figure 5). The revised cycle chart of Haq et al. [1987] shows many more events than the earlier Exxon publications; however, Haq et al. report six major supercycle boundaries for the Cenozoic which are the same as the major events of Vail et al. [1977]. For this study, the ages of these events were estimated using planktonic foraminiferal zonations given by Vail et al. [1977], Vail and Hardenbol [1979], and Vail and Mitchum [1980] recalibrated to the Berggren et al. [1985] time scale. The chronostratigraphy was derived from studies of the U.S. east coast margin, Irish margin, and northwest European epicontinental sea [Aubry, 1985; Melillo, 1985; Miller et al., 1985a,c; Poag et al., 1985; Snyder and Waters, 1985; Miller and Hart, 1987; Olsson and Wise, 1987; Poag et al., 1987]. Chronostratigraphic breaks and offlap events occurred near the middle/late Miocene, early/late Oligocene, Eocene/Oligocene, middle/late Eocene, early/middle Eocene, and early/late Paleocene boundaries (Figure 5). Further seismic stratigraphic evidence exists for erosion near the early/late Oligocene boundary [Miller et al., 1985c] and the middle/late Miocene boundary [Farre, 1985] in the form of canyons incised into continental margins. Similar timing of erosion on margins with different tectonic and sedimentologic histories suggests a global cause: eustatic change. We agree with the conclusions of Exxon researchers on this point; however, the record of erosion on passive continental margins is more complicated than initially described by Vail et al. [1977], as will be pointed out below.

There is a good correspondence between the timing of erosional events and the

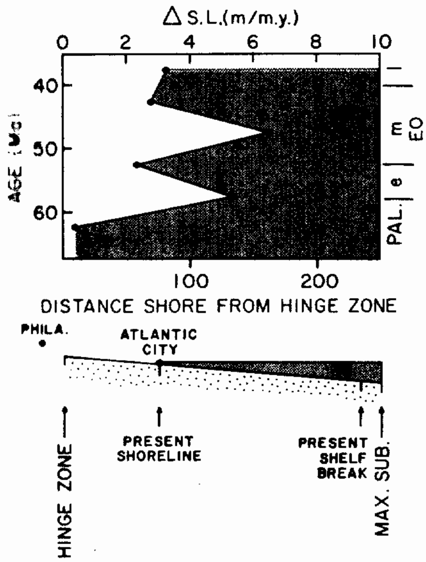

Fig. 6. Model for effect of Paleocene to Eocene eustatic changes on the passively subsiding U.S. east coast continental margin. (top) Rate of change of sea level (S.L.) in meters per million years estimated in 5-m.y. increments using the tectonoeustatic record of Kominz [1984]. (bottom) Section from the hinge zone (taken as near outcrop belt) to the point of maximum subsidence (taken as immediately seaward of present-day shelf break). Figure is modified after Pitman [1978]. By assuming maximum subsidence rate for the margin of 10 m/m.y. [Thorne and Watts, 1984], the area of the margin below sea level at a given time is equivalent to the shaded area at top.

$\delta^{18}O$ events (Figure 5), given uncertainties in dating unconformities (typically 1-2 m.y.). Two notable exceptions occur, one in the middle Miocene and one in the Paleocene. Oxygen isotope studies strongly suggest that a glacioeustatic event occurred in the middle Miocene (approximately 14 Ma). No major erosional event has been detected in the middle Miocene, but this is probably due to poor records. For example, coalesced unconformities have prevented the detection of a middle Miocene event on the U.S. Atlantic and Irish margins [Snyder and Waters, 1985; Miller and Hart, 1987]. Conversely, seismic sequence analyses suggest that there were one or more erosional events during the Paleocene, but no $\delta^{18}O$ increase has been noted. We propose that the lack of correlation of the $\delta^{18}O$ record with the Paleocene event(s) results from different mechanisms for sea level changes between the early and late Tertiary.

We suggest that the causal mechanism for major eustatic fluctuations of the past 36 m.y. was ice-volume change. There is no physical or isotopic evidence of glacioeustatic change prior to the Oligocene, and both isotopic and geological evidence indicate substantially ice-free conditions in the Paleocene to Eocene. For example, the lack of a $\delta^{18}O$ increase in planktonic foraminifera across the middle/late Eocene boundary has been suggested as indicating that this was not an ice growth event [Keigwin and Corliss, 1986]. Still, a major erosional event occurred at this time (Figure 5).

We suggest that continental margin erosion during the Paleocene and Eocene was caused by global seafloor spreading changes which changed the volume of the ocean basins and global sea level. Although such tectonoeustatic changes have been assumed to be far too slow to have caused major erosional events on passive margins [Pitman, 1978], the rate of eustatic lowerings caused by global spreading rate changes during the Paleocene and Eocene may be as high as 10 m/m.y. [Pitman, 1978; Pitman and Golovchenko, 1983; Kominz, 1984]. This rate is similar to the subsidence rate of old (>100 m.y.) passive continental margins [Thorne and Watts, 1984]. We therefore suggest that tectonoeustatic changes could severely affect passive continental margin sedimentation during the Paleogene. We illustrate (Figure 6) the Paleogene record of transgressions and regressions resulting from the interaction of tectonoeustatic changes and margin subsidence for an old margin: the U.S. east coast. We used the tectonoeustatic record of Kominz [1984], applying Pitman's [1978] model for a margin with maximum subsidence of 10 m/m.y. During Paleogene tectonoeustatic lowerings (e.g., about 57 Ma and 47 Ma in Figure 6), large areas of older margins were exposed subaerially (Figure 6), resulting in widespread development of unconformities in epicontinental seas and continental shelves. The material eroded from these exposed shelves would have increased sediment supply to the areally restricted submarine shelf, stimulating increased slope failure and submarine erosion (Figure 6).

The onlap-offlap record of Vail et al. (1977) implicitly suggests that the

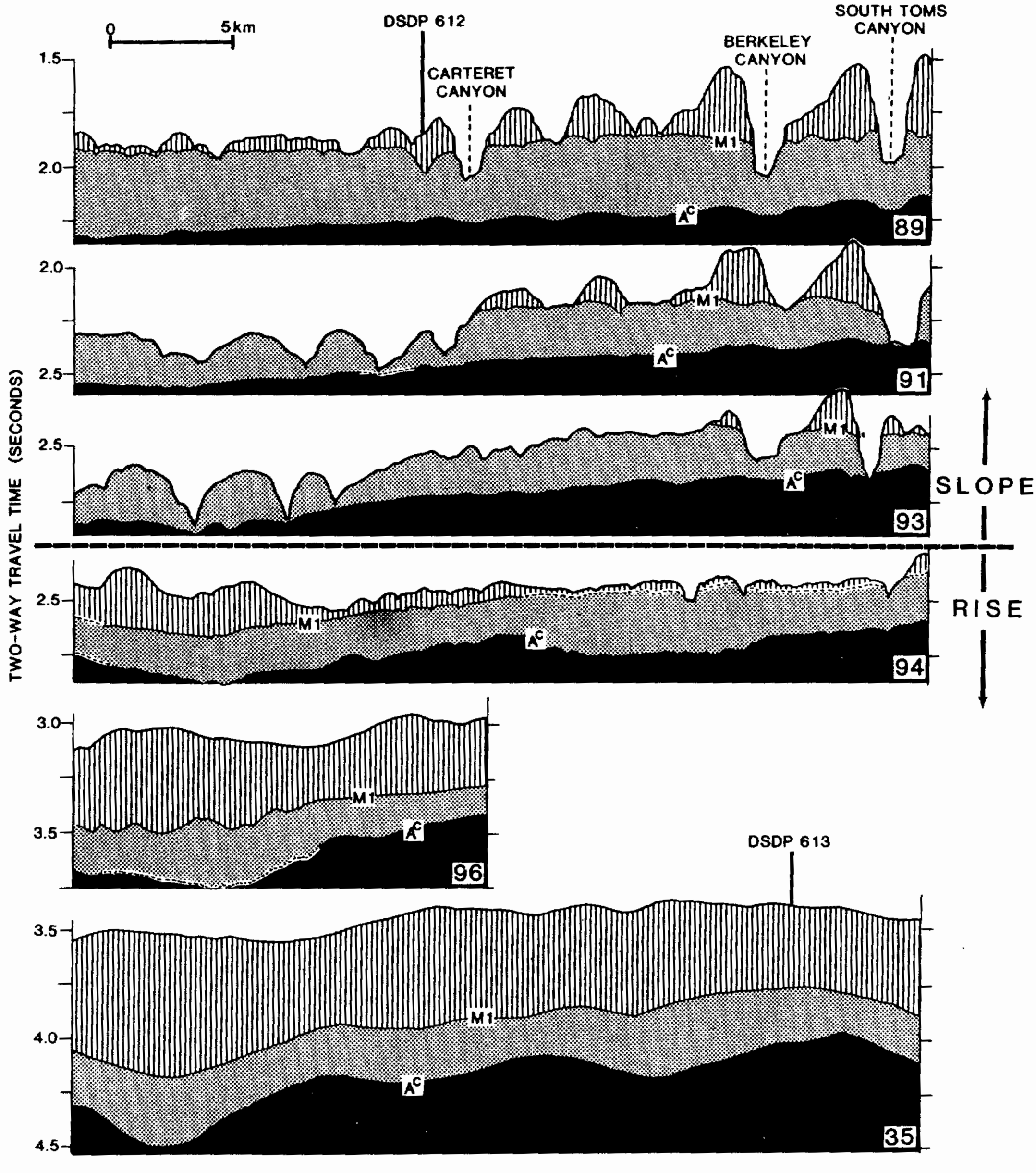

Fig. 7. Line tracings of high-resolution seismic strike lines [Robb, 1980] for the New Jersey continental slope and rise. Lines are arranged in order of increasing water depth. M1 denotes a reflector correlated with the middle/late Miocene boundary [K. G. Miller, A. J. Melillo, G. S. Mountain, and J. Farre, unpublished manuscript, 1987]; "Ac" is a reflector correlated with the lower/middle Eocene boundary. Note the V-shaped canyon associated with reflector M1 on line 89 near Site 612 which is similar to modern Carteret canyon. Note the U-shaped lower slope channels associated with the present-day seafloor, reflector M1, and reflector "Ac" on line 35.

mechanism for continental margin erosion has been the same throughout the Tertiary. However, because large continental ice sheets did not develop until the Oligocene, we suggest that mechanisms and rates of sea level change were different between the early and late Tertiary. We speculate that the contrasting mechanisms and rates of eustatic change between the early and late Tertiary may have caused different erosional patterns on passive continental margins.

Examination of seismic stratigraphic data from the New Jersey continental slope and rise supports the speculation that different erosional geometries occur on this margin. Post-Eocene erosion is associated with V-shaped canyons similar to Pleistocene canyons (Figure 7); the lower slope and upper rise expressions of these canyons are U-shaped channels filled with shallow-water (neritic) material, indicating breaching of the shelf break [Katz and Miller, 1987]. In contrast, V-shaped canyons have not been observed in Paleocene and Eocene seismic sections on the New Jersey margin (the Oligocene is not well represented on this margin because of coalesced unconformities [Miller et al., 1985c]). U-shaped lower slope and upper rise channels also have been observed in the Paleocene and Eocene, but in constrast to the post-Eocene channels, these Paleocene to Eocene channels were cut and filled with sediments derived from the lower slope and upper rise (Figure 7).

These apparently different erosional styles are consistent with different mechanisms of sea level change for the early and late Tertiary. During the Paleocene and Eocene, slow ocean basin volume changes (tectonoeustacy) caused partial exposure of shelves and slumping on the slope and rise. Rapid post-Eocene ice volume changes (glacioeustacy) exposed most of the shelves and eroded V-shaped canyons on the slope. Although tectonoeustatic changes occurred in the post-Eocene, the rate of tectonoeustatic change was slower than in the pre-Oligocene (less than 2 m/m.y. except for the interval 15-10 Ma [Kominz, 1984]), and sea level change was dominated by rapid, large glacioeustatic changes.

Since our seismic stratigraphic studies are limited to one margin, the different erosional style may be attributed to local effects (faulting, sedimentological changes, etc.). In fact, the change in erosional patterns on the New Jersey margin is associated with a change from carbonate to clastic regime [Poag, 1985], which could contribute to geometrical differences. Interregional comparisons of various margins with different thermal and sedimentological histories are needed to properly test the concept of different erosional geometries between the early Tertiary and late Tertiary. We believe that such comparison will establish that the differences observed are not peculiarities of New Jersey margin but are fundamentally different responses of continental margins to different mechanisms of sea level change.

CONCLUSIONS

1. On the basis of stable isotope and geological evidence, we suggest that the world was probably ice free throughout most of the Paleocene and Eocene.
2. Benthic foraminiferal $\delta^{18}O$ values indicate that significant continental ice sheets have existed since the beginning of the Oligocene (approximately 36-35 Ma), although ice sheets may have disappeared during portions of the Oligocene and early Miocene. Covariance of benthic and low- to middle- latitude planktonic $\delta^{18}O$ records suggests several ice growth and decay events with growth at ca. 35, 31, 25, 14, and 10 Ma.
3. Glacioeustatic changes resulted in post-Eocene continental margin erosion; global spreading rate changes caused tectonoeustatic lowerings which resulted in pre-Oligocene erosion.
4. These contrasting mechanisms of eustatic changes between the early and late Tertiary may have caused different erosional geometries, although this needs to be tested with interregional comparisons of various margins.

Portions of the Cenozoic $\delta^{18}O$ record require better documentation to determine the nature and timing of apparently important, but poorly understood changes. In particular, we look toward continued documentation of the events noted in the middle to late Eocene (approximately 52 and 41 Ma), Oligocene (approximately 31 and 25 Ma), and early late Miocene (approximately 10 Ma) in the detail of the Eocene/Oligocene boundary and middle Miocene events.

Acknowledgments. We thank W. B. Curry, L. D. Burckle, J. Farre, T. R. Janecek,

D. V. Kent, D. Martinson, A. J. Melillo, M. L. Prentice, and W. B. Ruddiman for discussions, L. D. Keigwin, J. P. Kennett, H. Oberhansli, S. M. Savin, and E. Vincent for supplying data while in press, J. P. Kennett, N. J. Shackleton and S. M. Savin for reviewing the submitted mauscript, and M. E. Katz and D. King for technical assistance. Samples were provided by the DSDP. Supported by NSF Grants OCE85-00859 (KGM), OCE82-08784 and OCE84-02055 (RGF) and by a grant from the Arco Foundation. L-DGO contribution 4090.

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(Received February 7, 1986;
revised December 1, 1986;
accepted December 1,1986.)