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CENOZOIC DEEP-SEA CIRCULATION: EVIDENCE FROM DEEP-SEA BENTHIC FORAMINIFERA

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Deep-sea benthic foraminiferal faunas reflect the deep oceanic environment, the character of which is determined by interaction of deepwater circulation patterns, physicochemical parameters of the surface waters in the deepwater source areas, and nutrient influx from primary productivity in overlying surface waters. Three periods of turnover in deep-sea benthic foraminiferal assemblages can be recognized in Cenozoic sequences: (1) rapid ($<10^4$ yr), global extinction in the latest Paleocene, followed by migration and diversification; (2) gradual turnover in the late middle Eocene through early Oligocene, characterized by a decrease in diversity, a decrease in relative abundance of *Nuttallides truempyi* followed by its extinction, and a decreasing relative abundance or disappearance of *Bulimina* species in the lower bathyal to upper abyssal zones; and (3) gradual turnover in the late early through middle Miocene, characterized by the decrease in relative abundance or disappearance of uniserial species from the lower bathyal to abyssal reaches, the migration of miliolid species into these regions, and the evolution of *Cibicidoides wuellerstorfi*. The rapid mass extinction (35–50% of species) of deep-sea benthic foraminifera in the latest Paleocene was coeval with a transient 1–2‰ decrease in oxygen and carbon isotope ratios in benthic as well as planktonic foraminifera, superimposed on longer-term changes. The extinction could have resulted from a shift in dominant deepwater formation from high to low latitudes. Such a shift would change temperature and oxygen content of the intermediate to deep waters, but it would also change local nutrient input by changing global patterns of upwelling of nutrient-rich waters to the surface and thus of high-productivity areas. Faunal evidence suggests that this "reversed" pattern of oceanic circulation persisted no longer than the early Eocene, and possibly not more than about half a million years. The two periods of gradual benthic faunal changes overlap in time with two relatively rapid (of the order of 10^5 years) shifts toward heavier oxygen isotopic values of benthic foraminifera, in the earliest Oligocene and middle Miocene. Faunal changes started before the isotopic changes and were more gradual. The faunal changes might reflect periods of gradual change in the physicochemical character of surface waters in the source areas of deepwater formation (e.g., decrease in temperature), as well as changes in oceanic productivity. The more rapid changes in oxygen isotopic values are not directly reflected in benthic foraminiferal assemblage changes and might represent, at least in part, a rapid buildup of ice volume on land, a process that cannot be reflected in the benthic foraminifera faunas.

INTRODUCTION

The deep oceanic environment has been more stable through geologic time than the surface environments of the Earth, with less short-term variability, although spatial "patchyness" occurs [*Thiel et al.*, 1988; *Gooday and Lambshead*, 1989; *Lambshead and Gooday*, 1990]. This stability is reflected in the slow faunal turnover of benthic foraminiferal faunas as compared to planktonic microfossils. Benthic faunas have been a relatively neglected group and are commonly overlooked in reconstructions of deepwater circulation patterns of the past. Much information on Cenozoic deep-sea faunas has been obtained as an outcome of recent Ocean Drilling Program (ODP) drilling operations in the Indian and Southern oceans (legs 113, 114, 118, 119, and 121). The interpretation of these data is not straightforward, because no unequivocal relations between benthic foraminiferal faunal composition and environmental parameters have been discovered, and there exists no transfer function to translate benthic foraminiferal data into environmental parameters. The collection of long-term records, however, has helped to delineate benthic foraminiferal faunal history, making it possible to delineate periods of faunal change, alternating with long, relatively stable periods. Benthic foraminiferal faunal Cenozoic history proceeded stepwise, as did ocean history [e.g., *Berger et al.*, 1981; *Barron and Baldauf*, 1989].

Dramatic changes in the Cenozoic oceans are documented in stable isotopic and biotic records [e.g., *Savin et al.*, 1981; *Shackleton and Kennett*, 1975; *Kennett and Shackleton*, 1976; *Kennett*, 1977; *Savin*, 1977; *Berger*, 1981; *Berger et al.*, 1981; *Kemp*, 1978; *Shackleton and*

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Boersma, 1981; *Mercer*, 1982; *Shackleton*, 1986; *Miller et al.*, 1987*a*; *Kennett and Barker*, 1990; *Thomas*, 1989, 1990*a*; *Rea et al.*, 1990; *Webb*, 1990; *Barron et al.*, 1991; *Wise et al.*, 1992; *Kennett and Stott*, 1990, 1991; *Zachos et al.*, 1992*c*; *McGowran*, 1989, 1991]. Similarly dramatic changes were recorded in land floras [*Wolfe*, 1978; *Wolfe and Poore*, 1982; *Wing*, 1984; *Schmidt*, 1991; *Wing et al.*, 1991], as the Earth changed from essentially unglaciated to a world with very cold, glaciated polar regions.

Climatic change as reflected in the oxygen and carbon isotopic records did not proceed gradually, but stepwise [e.g., *Berger et al.*, 1981]. Deep waters in the world's oceans and surface waters at high latitudes cooled strongly after the very warm late Paleocene to earliest Eocene, the warmest period of the Cenozoic. Early Eocene surface water temperatures at high latitudes (65°S) are estimated to have been about 15°–17°C [*Stott et al.*, 1990; *Kennett and Stott*, 1991; *Wise et al.*, 1992; *Zachos et al.*, 1992*a*, *c*]. Climate as well as deepwater circulation and ocean chemistry during the Paleocene strongly resembled the Late Cretaceous situation [*Thomas*, 1990*b*; *Corfield et al.*, 1991]. At some time during the middle Eocene through early Oligocene the psychrosphere was established [*Benson*, 1975], as well as at least partial ice sheets on eastern Antarctica [*Savin et al.*, 1981; *Kennett and Shackleton*, 1976; *Keigwin and Keller*, 1984; *Miller and Thomas*, 1985; *Miller et al.*, 1987*a*; *Kennett and Barker*, 1990; *Barron et al.*, 1991; *Spezzaferri and Premoli-Silva*, 1991; *Wise et al.*, 1992]. The extent and nature of these ice sheets (true continental ice sheets, temperate ice sheets, or upland and coastal glaciers) are under intensive discussion [*Kennett and Barker*, 1990; *Barron et al.*, 1991; *Wise et al.*, 1992; *Zachos et al.*, 1992*a*, *b*, *c*]. Evidence from glaciomarine sediments in the Prydz Bay area (East Antarctica) suggests that some Antarctic ice sheets or large glaciers reached sea level during the earliest Oligocene, at least temporarily [*Barron et al.*, 1991], in agreement with the observation of ice-rafted material in lowermost Oligocene sediments on the Kerguelen Plateau [*Wise et al.*, 1992; *Zachos et al.*, 1992*a*, *b*, *c*].

There is no agreement on the interpretation of the Cenozoic oxygen isotopic record of benthic foraminifers, especially in how far it demonstrates establishment and growth of ice sheets (the ice volume effect) and in how far it represents cooling of deep waters in the oceans [e.g., *Matthews and Poore*, 1980; *Poore and Matthews*, 1984; *Keigwin and Corliss*, 1986; *Shackleton*, 1986; *Miller et al.*, 1987*a*; *Prentice and Matthews*, 1988; *Wise et al.*, 1992; *Zachos et al.*, 1992*a*, *c*; *Oberhänsli et al.*, 1991; *Spezzaferri and Premoli-Silva*, 1991]. Additional complications in the benthic record might result from storage of the heavier oxygen isotopes in deeper waters when these are formed at least partially by evaporation [*Railsback et al.*, 1989; *Railsback*, 1990].

It is not clear whether equatorial surface waters were cooler than at present during some globally warm periods [e.g., *Shackleton*, 1984, 1986] or remained essentially at the same temperature throughout the Cenozoic, as indicated by the distribution of tropical biota such as hermatypic corals, mangroves, and larger foraminifera [*Matthews and Poore*, 1980; *Adams et al.*, 1990; *McGowran*, 1989, 1991]. Especially during the Oligocene there is a discrepancy between macrofaunal and macrofloral, as well as microfaunal, data indicating fairly warm climates at high to middle latitudes and isotopic data suggesting cooler climates [*Adams et al.*, 1990; *McGowran*, 1991; *Spezzaferri and Premoli-Silva*, 1991]. There is no agreement on whether the flat Eocene-Oligocene latitudinal oxygen isotopic gradient in planktonic foraminifera truly reflects a very low temperature gradient in surface waters [*Shackleton and Boersma*, 1981; *Premoli-Silva and Boersma*, 1984; *Shackleton*, 1984; *Keigwin and Corliss*, 1986; *Boersma et al.*, 1987; *Boersma and Premoli-Silva*, 1991; *Zachos et al.*, 1992*a*]. The distribution of planktonic biota suggests a steeper temperature gradient than the oxygen isotopic data [*Wei*, 1991], and the planktonic oxygen isotope record may be complicated by local influx of fresh water into high-latitude surface waters [e.g., *Wise et al.*, 1991, 1992; *Zachos et al.*, 1992*c*].

There is similar discussion on modes of deepwater formation in the Cenozoic. The question is whether intermediate and bottom waters dominantly formed at high latitudes as a result of an increase in density because of cooling [e.g., *Manabe and Bryan*, 1985; *Barrera et al.*, 1987; *Katz and Miller*, 1991; *Thomas*, 1989, 1990*a*, *b*] or at low latitudes after an increase in density by evaporation [*Chamberlin*, 1906; *Matthews and Poore*, 1980; *Brass et al.*, 1982; *Hay*, 1989; *Woodruff and Savin*, 1989; *Kennett and Stott*, 1990]. Oceanic circulation dominated by deep and/or intermediate waters originating from low-latitude sources represents a reversal of the present deep-oceanic circulation [e.g., *Keith*, 1982; *Prentice and Matthews*, 1988; *Woodruff and Savin*, 1989; *Kennett and Stott*, 1990]. Oceans "running the reverse" from the modern circulation pattern might be required to model satisfactorily the high heat transfer from low to high latitudes required to maintain the warm Eocene climate at high latitudes [*Barron*, 1985, 1987; *Barron and Peterson*, 1991]. Carbon and oxygen isotopic data as well as deep-sea benthic foraminiferal data, however, may be seen as indicative of a Late Cretaceous–Paleocene Atlantic and Pacific circulation dominated by deep and intermediate waters formed at high latitudes, with possible exception of one or more short (<0.5 m.y.) periods of high-volume formation of warm, salty bottom waters during the early to early middle Eocene [*Barrera et al.*, 1987; *Miller et al.*, 1987*b*; *Thomas*, 1989, 1990*a*; *Katz and Miller*, 1991; *Zachos et al.*, 1992*c*; *Pak et al.*, 1991; *Barron and Peterson*, 1991; *Pak and Miller*, 1992]. The Indian

Ocean might have contained a relatively large volume of waters from a low latitude, as a result of outflow from the eastern end of the Tethys Ocean until its closure in the middle Miocene [*Woodruff and Savin*, 1989; *Nomura*, 1991; *Zachos et al.*, 1992*c*].

Deep waters formed at high latitudes during the periods in which the Antarctic was essentially ice free would have had a lower oxygen content than the present deep waters, because of the higher temperatures of the surface waters in the source areas, and thus lower oxygen content at the time of formation. The oxygen content would have been even lower in waters formed by evaporation at low latitudes, because of the nonlinear, reverse relation between solubility of oxygen and temperature. The ratio of the solubility of O_2 at 0°C to that at 24°C is 1.6 (solubilities taken per atmosphere pressure of the gas) [*Broecker and Peng*, 1982]. In recent oceans, most surface waters are supersaturated in oxygen by a few percent; the oxygen solubilities range from about 190 μmol/kg at 30°C, 225 mmol/kg at 20°C, 275 μmol/kg at 10°C, to 350 μmol/kg at 0°C. In the northern Pacific the oxygen content of the deep waters is the lowest, because these waters have been out of contact with the oxygen supply in the atmosphere for more than 1000 years. In this region the apparent oxygen utilization (AOU, the difference between saturated oxygen content and observed oxygen content) is about 190 μmol/kg [*Broecker and Peng*, 1982]. This value represents the amount of oxygen used during the travel of the deep waters from their source area, as a result of decay of organic material and respiration by the bottom-dwelling fauna. If the waters in the source area had been at a temperature of 30°C (and would have been dense enough to sink to the deep oceans at that temperature), they could not have taken up more than 190 μmol/kg. The oxygen in these waters would have been exhausted by the time that they had been out of contact with the atmosphere for about 1000 years (unless the oxygen content of the atmosphere was considerably higher at the time). Thus the warmer deep waters of the past possibly contained less dissolved oxygen than the present deep waters. A more sluggish deepwater ventilation would decrease the dissolved oxygen content even more [e.g., *Thierstein*, 1989]. Chemical and circulation modeling of the oceans suggests that dominance of warm saline deep waters in the oceans might likely drive the oceans to anoxia, which did not occur during the last 90 million years [*Herbert and Sarmiento*, 1991]. Benthic foraminiferal faunas have been interpreted as indicating good deep-ocean ventilation in Late Cretaceous through Paleocene and middle Eocene and later [*Thomas*, 1990*a*; *Kaiho*, 1991].

The ratio of the solubility of CO_2 at 0°C to that at 24°C is 2.2 (solubilities taken per atmosphere pressure of the gas [*Broecker and Peng*, 1982]). The amount of CO_2 that can be dissolved in the colder waters of the present oceans is thus much larger than the amount that could dissolve in the warmer oceans of the past. At lower temperatures the solubility of calcite is much higher than at high temperatures, so that the cold, present-day deep waters are potentially more corrosive to $CaCO_3$. More vigorous deepwater ventilation, however, could counteract this effect, because of the presence of overall "younger" deep waters (out of contact with the atmosphere for a shorter period) in an ocean with a faster turnover rate.

High-resolution carbon and oxygen isotopic records from many locations in the oceans and high-resolution, reliable stratigraphic data are necessary to develop models of deepwater sources, especially how many significant source areas there were, what their relative contribution to the total deepwater volume was, and where they were [e.g., *Woodruff and Savin*, 1989; *Mead and Hodell*, 1992; *Pak and Miller*, 1992]. We should not forget that all isotope information is derived from sites above the calcium carbonate compensation depth (CCD), and thus we do not know what characterized water masses in the middle and lower abyssal realms.

Recent deepwater benthic foraminiferal faunas reflect the complex interaction of deep oceanic circulation, the character of the surface waters in the source regions, and local primary productivity influx [e.g., *Lohmann*, 1978; *Douglas and Woodruff*, 1981; *Lutze and Coulbourn*, 1984; *Culver*, 1987; *Gooday*, 1988; *Gooday and Lambshead*, 1989]. Benthic foraminifera may form as much as 50% or more of the eukaryotic biomass in the deep sea [*Gooday et al.*, 1992]. They are the only fossil-providing organisms that live in large enough numbers on the nutrient-starved ocean floor environment to be represented in Deep Sea Drilling Project (DSDP)–ODP sized core samples in numbers large enough for statistically valid studies, at high time resolution. Therefore data on benthic foraminiferal faunal composition should be used as constraints in the reconstruction of deepwater formational processes as inferred from stable isotope or trace element studies. The large influence of nutrient influx on the deep-sea benthic foraminiferal faunas should be kept in mind, while assessing the influence of deepwater circulation on faunal patterns [*Gooday and Lambshead*, 1989; *Gooday and Turley*, 1990; *Lambshead and Gooday*, 1990].

In this paper I review information that has recently become available as a result of ODP drilling in the Southern and Indian oceans and interpret its significance for ocean circulation models for the Cenozoic, with emphasis on the Paleogene. The bathymetric division follows *Berggren and Miller* [1989] and *van Morkhoven* [1986]: neritic, <200 m; upper bathyal, 200–600 m; middle bathyal, 600–1000 m; lower bathyal, 1000–2000 m; upper abyssal, 2000–3000 m; lower abyssal, >3000 m. The information presented in this paper is dominantly based on calcareous taxa of benthic foraminifera. The calcium carbonate compensation depth during the Cenozoic fluctuated between 3500 and 5000

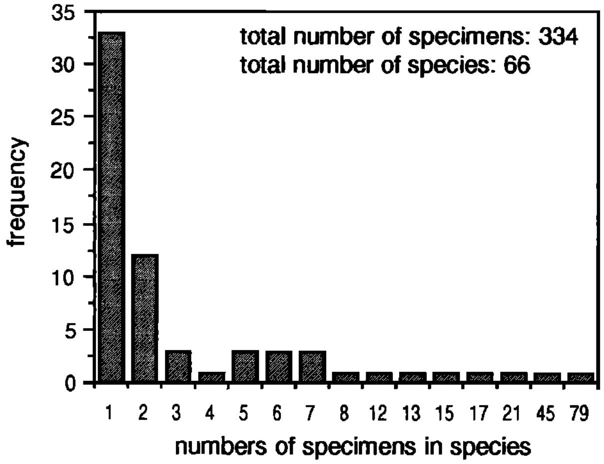

Fig. 1. Distribution of specimens within species for a highly diverse, Paleocene sample of benthic foraminifera (Site 690). Note the discontinuity in scale along the horizontal axis.

m [*van Andel*, 1975]; thus this information does not extend to the deeper parts of the ocean basins. The time scale used throughout is after *Berggren et al.* [1985], except where it has been modified by *Aubry et al.* [1988]. Major changes in numerical ages (especially in the Paleogene) will probably be proposed in the near future [*Montanari*, 1990; I. Premoli-Silva, personal communication, 1991; S. C. Cande and D. V. Kent, work in preparation] but are not yet available.

DEEP-SEA BENTHIC FORAMINIFERA AND PALEOCEANOGRAPHY

Deep-sea benthic foraminifera are not easy subjects of study. In most deep-sea samples, with the exception of those strongly affected by dissolution, they are outnumbered by several orders of magnitude by the tests of planktonic organisms, so that it is time consuming to extract enough specimens for a statistically valid study. This is aggravated by the fact that deep-sea benthic foraminifera form highly diverse assemblages, as do other deep-sea organisms, and consist of specimen-poor, species-rich assemblages [e.g., *Sanders*, 1968; *Douglas and Woodruff*, 1981]. Large numbers of specimens must be studied to obtain a valid representation of the total species richness, and many of the total number of species present are represented by only one or two specimens (Figure 1; *Douglas and Woodruff* [1981]). The minimum number needed to represent the species richness depends upon the diversity and can be determined by plotting rarefaction curves (*Sanders* [1968]; see also *Thomas* [1985]). The stratigraphic ranges of the rare species cannot be determined with precision [e.g., *Signor and Lipps*, 1982]. Biology of recent deep-sea benthic foraminifera is not well known; they feed at low trophic levels (consuming planktonic debris and bacteria), and some species can respond quickly to the pulsed influx of detritus [*Gooday et al.*, 1992].

As a result of the high diversity and the presence of many rare species, results from quantitative studies are difficult to present. Data are commonly presented after *Q* mode or *R* mode multivariate analysis, or cluster analysis, on relative and/or absolute abundance data, and values of derived factors are plotted to supplement the information shown in simple species' relative or absolute abundances [e.g., *Lohmann*, 1978]. The derived factors, however, are commonly difficult to interpret unequivocally. In different studies, factors have been interpreted as representing water mass properties [*Lohmann*, 1978; *Bremer and Lohmann*, 1982], a combination of water mass properties, substrate and nutrient influx [*Lutze and Coulbourn*, 1984], or other combinations of environmental parameters. Nutrient influx in the shape of aggregates of phytodetritus appears to be an important factor in species composition as well as absolute abundance of recent deep-sea faunas [*Gooday*, 1988; *Gooday and Lambshead*, 1989; *Gooday et al.*, 1992]. Evaluation of the importance of this parameter for the past (for example, by using organic carbon content of the sediment as a proxy for nutrient influx) indicates that nutrient influx is indeed important [e.g., *Caralp*, 1984; *Lutze and Coulbourn*, 1984]. These studies, however, have not resulted in a transfer function relating absolute or relative abundances of species to nutrient influx.

In stratigraphic studies the factors commonly do not show a simple vertical succession, but several assemblages (as indicated by high loadings on a factor) alternate vertically in a drill hole or stratigraphic section. These assemblages thus probably represent not evolutionary and extinction events, but migratory events resulting from reversible environmental changes [e.g., *Mueller-Merz and Oberhänsli*, 1991; *Nomura*, 1991*a*; *Oberhänsli et al.*, 1991; *Mackensen*, 1992].

In few instances the relative abundance of species has been related to specific environmental properties. The recent species *Nuttallides umbonifera*, for instance, has been reported to be most abundant in waters that are highly corrosive to $CaCO_3$, thus with high concentrations of dissolved CO_2 [*Bremer and Lohmann*, 1982]. *Eilohedra weddellensis*, *Epistominella exigua*, *Cassidulina teretis*, and *Melonis barleeanus* appear to be dominant in faunas where there is a high influx of phytodetritus [*Caralp*, 1984; *Gooday*, 1988; *Gooday and Lambshead*, 1989; *Gooday et al.*, 1992]. Several biserial and triserial taxa such as *Bolivina* and *Uvigerina* spp. and thin-walled species of *Cassidulina* have been well described as abundant to dominant in low-O_2 environments, but their abundance may be primarily controlled by high nutrient as well as low O_2 conditions [e.g., *Douglas*, 1981; *Sen Gupta et al.*, 1981; *Caralp*, 1984; *Lutze and Coulbourn*, 1984; *Bernard*, 1986; *Corliss and*

Chen, 1988; *Niensted and Arnold*, 1988; *van der Zwaan et al.*, 1992].

Recently, attempts have been made to correlate not abundances of species, but general morphological types with environmental parameters, especially to an infaunal or epifaunal mode of life [*Corliss*, 1985; *Corliss and Chen*, 1988; *Corliss and Emerson*, 1990; *Rosoff and Corliss*, 1992]. These authors concluded that high relative abundance of infaunal morphotypes (including the biserial and triserial groups) reflects a relatively high flux of nutrients to the ocean floor. A similar approach for fossil material has been tried [*Keller*, 1988; *Thomas*, 1989, 1990*a*, *b*; *Kaiho*, 1991; *Oberhänsli et al.*, 1991], but the validity of the correlation between test morphology and life style is not beyond doubt.

The existing data base on the geological record of deep-sea benthic foraminiferal faunas is thus commonly difficult to access for nonmicropaleontologists; there is no simple numerical parameter to be plotted representing faunal composition, and there is no globally valid biostratigraphic zonation, or environmental zonation, or depth zonation. As a result, this data base is usually not consulted by nonmicropaleontologists while reconstructing circulation patterns and deepwater physicochemical properties. As an example, *Keith* [1982] argued that the end-Cretaceous extinction episode was caused by anoxia in the deep oceans, followed by catastrophic overturns and extinction of planktonic organisms as a result of poisoning. This theory conflicts with deep-sea benthic foraminiferal evidence, because it has long been known that extinction of deep-sea benthic foraminifera at the end of the Cretaceous was not catastrophic and hardly reached above background levels [*Cushman*, 1946; *Beckman*, 1960; *Webb*, 1973; *Dailey*, 1983; *Douglas and Woodruff*, 1981].

A biostratigraphic zonation for deep-sea benthic foraminifera (bathyal and abyssal) was proposed by *Berggren and Miller* [1989] (Figure 2), largely using material from Atlantic DSDP sites or land sections in the Caribbean region. Many of the zonal species, however, do not occur or are very rare at other sites or have different ranges. The timing of zonal boundaries, on the other hand, and thus of periods of faunal change, appears to be recognizable over wider areas, especially for the Paleogene. Faunal changes as recognized by the first and last appearances appeared to cluster around the times of zonal boundaries of *Berggren and Miller* [1989] for lower bathyal to uppermost abyssal faunas from Maud Rise, Antarctica [*Thomas*, 1990*a*] (Figure 2).

Informal benthic zones based upon cluster analysis of quantitative faunal data were proposed for parts of the Cenozoic for lower bathyal to upper abyssal faunas from the Walvis Ridge (DSDP Site 525), where an unconformity is present between middle Eocene and upper Oligocene [*Boltovskoy and Boltovskoy*, 1989] (Figure 2). For middle to upper bathyal sites on Ninetyeast Ridge and Broken Ridge (Indian Ocean), fewer assemblages were recognized using multivariate analysis [*Nomura*, 1991*a*, *b*] (Figure 2), but periods of faunal change were coeval with zonal boundaries in the work of *Berggren and Miller* [1985]. In sections from Japan and New Zealand and material from DSDP sites in the North Pacific, the South Atlantic, and the Indian Ocean, *Kaiho* [1991, 1992] recognized four benthic zones, with boundaries overlapping in time with some of the zonal boundaries of *Berggren and Miller* [1989]. His data have low time resolution (less than one data point per million years) and are gathered at many different sites, so that records had to be spliced, with the resultant problems of precision in time correlation. He concludes that these faunal changes might reflect different oxygenation states of the oceanic bottom waters and that there was relatively low oxygenation of deep waters in the early to earliest middle Eocene (in agreement with *Thomas* [1989, 1990*a*]).

In the southernmost Atlantic east of the Falkland Plateau, *Katz and Miller* [1991] recognized the major faunal change close to the end of the Paleocene, which is prominent in all zonal schemes in Figure 2, in addition to faunal overturn over a wide depth range (1000–2500 m) at 54 Ma. At the southern tip of Kerguelen Plateau (Indian Ocean) faunal assemblages in the lower bathyal range changed in the late middle Eocene (44 Ma), in the middle late Eocene (39 Ma), shortly after the end of the Eocene (36 Ma), and in the early Oligocene (32 Ma) [*Schroeder-Adams*, 1991]. At middle bathyal to uppermost abyssal depths on the northern Kerguelen Plateau (Indian Ocean), major faunal change as demonstrated in multivariate analysis of quantitative faunal data occurred in the last part of the Paleocene and at the middle to late Eocene boundary [*Mackensen and Berggren*, 1992].

These results appear to be roughly in agreement, but in detail many differences in timing are seen (Figure 2), which can only partially be attributed to problems in stratigraphic correlation. Many of the data sets have fairly high resolution, and isotope, biostratigraphic, and magnetostratigraphic data allow reasonably reliable correlation. There is considerable agreement in the different evaluations of deep-sea benthic foraminiferal data if one realizes that some of the apparently conflicting data on the timing of zonal boundaries or assemblage ranges result from the difficulty of assigning an exact location to a boundary, when faunal change occurs gradually over one or several millions of years. Many authors have recognized that there was an extended period of faunal change over the full middle bathyal to abyssal depth range, starting in the late early Miocene through the middle to late middle Miocene [*Woodruff and Douglas*, 1981; *Thomas*, 1985, 1986*a*, *b*; *Woodruff*, 1985; *Schnitker*, 1979*b*, 1986; *Murray et al.*, 1986; *van Morkhoven et al.*, 1986; *Miller and Katz*, 1987; *Boltovskoy and Boltovskoy*, 1988, 1989; *Nomura et al.*, 1991*b*, 1992]. Another extended period of faunal

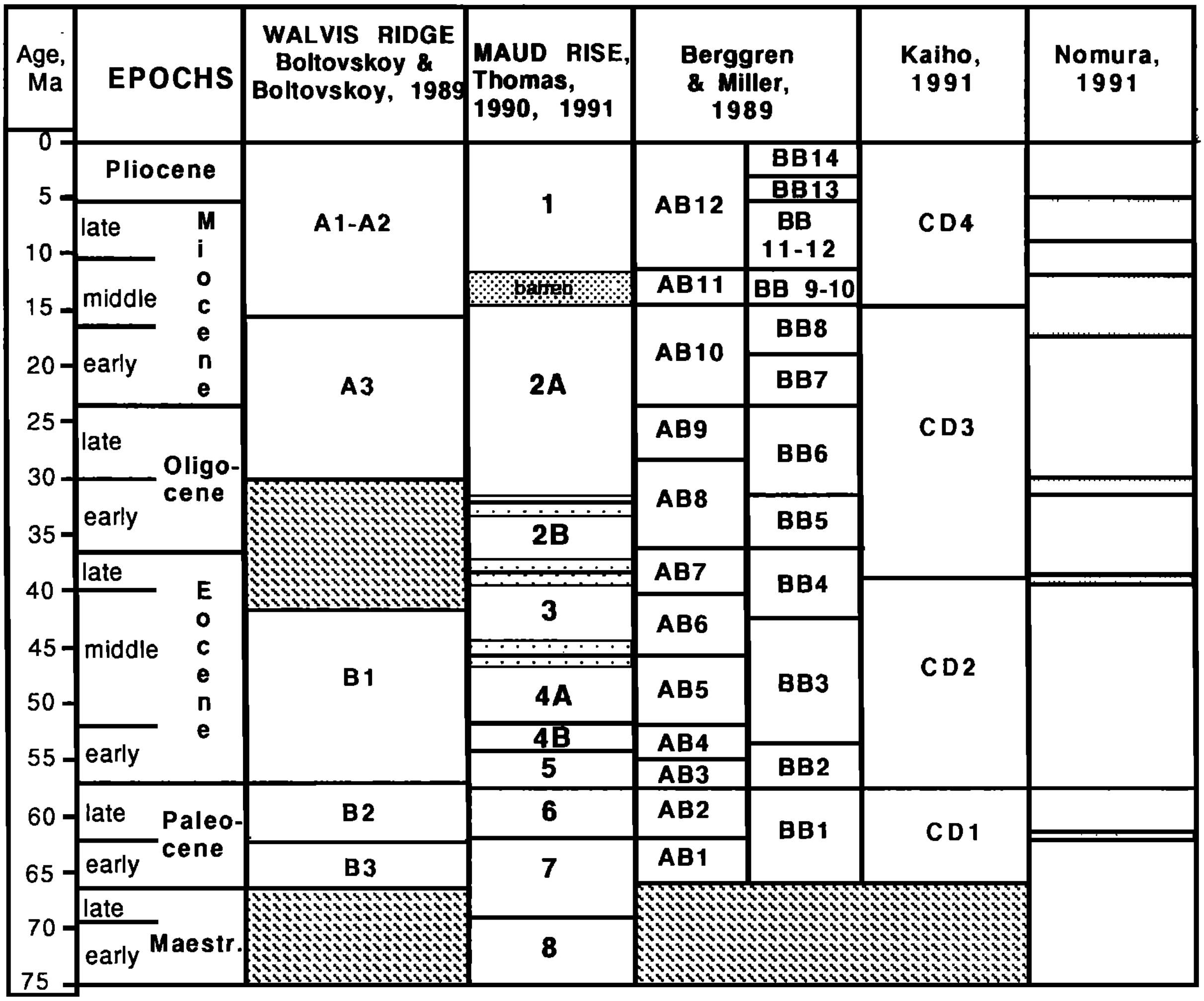

Fig. 2. Compilation of assemblages and informal and formal biostratigraphic zones of deep-sea benthic foraminifera. Data after *Boltovskoy and Boltovskoy* [1989], *Thomas* [1990*a*], *Berggren and Miller* [1989], *Kaiho* [1991], and *Nomura* [1991*a*, *b*]. Numerical ages after *Berggren et al.* [1985], except where amended by *Aubry et al.* [1988].

change started toward the end of the middle Eocene and ended close to the early/late Oligocene boundary [e.g., *Corliss*, 1981; *Tjalsma and Lohmann*, 1983; *Miller*, 1983; *Miller et al.*, 1984; *Boersma*, 1984, 1985, 1986; *Corliss and Keigwin*, 1986; *McGowran*, 1987; *Berggren and Miller*, 1989; *Boltovskoy*, 1980; *Boltovskoy and Boltovskoy*, 1988, 1989; *Mueller-Merz and Oberhänsli*, 1991; *Oberhänsli et al.*, 1991; *Thomas*, 1992] (see reviews in the work of *Douglas and Woodruff* [1981] and *Culver* [1987]).

In contrast with these extended periods of faunal change was the most profound benthic faunal change over the last 75 m.y. that occurred at the end of the Paleocene [*Cushman*, 1946; *Braga et al.*, 1975; *Schnitker*, 1979*b*; *Tjalsma and Lohmann*, 1983; *Miller et al.*, 1987*b*; *Berggren and Miller*, 1989; *Boltovskoy and Boltovskoy*, 1988, 1989; *Katz and Miller*, 1991; *Mackensen and Berggren*, 1992; *Nomura*, 1991; *Kaiho*, 1988, 1991; *Thomas*, 1989, 1990*a*, *b*; *Katz and Miller*, 1991; *Berggren et al.*, 1992]. This was the only benthic foraminiferal mass extinction documented, resulting in a loss of diversity of 35 to 50% (Figure 2). The extinction occurred over 10,000 years or less [*Thomas*, 1990*b*, 1991; *Kennett and Stott*, 1991; E. Thomas and N. J. Shackleton, work in preparation]. At many sites, another, less prominent period of faunal change was observed at the boundary between upper and lower Paleocene (Figure 2); over this period, however, there are not many high-resolution data sets available. There was no major deep-sea benthic foraminiferal extinction at the end of the Cretaceous, and Paleocene faunas closely resemble upper Maestrichtian faunas [*Beckmann*, 1960; *Dailey*, 1983; *Keller*, 1988; *Thomas*, 1990*b*; *Widmark*, 1990; *Widmark and Malmgren*, 1992; *Kaiho*, 1992].

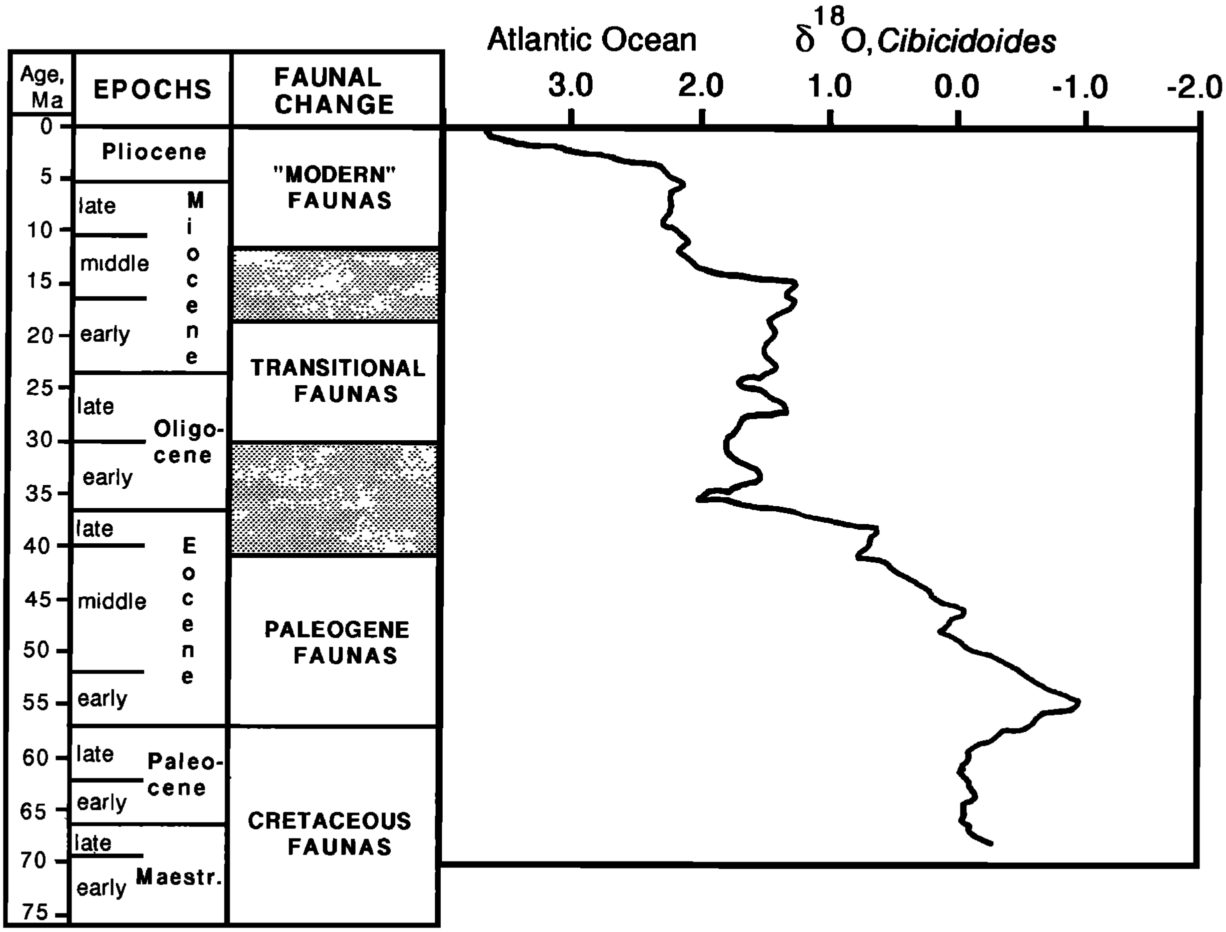

Fig. 3. Simplification of deep-sea benthic foraminiferal data, compared with the combined benthic foraminiferal oxygen isotopic curve for the Atlantic Ocean [*Miller et al.*, 1987*a*]. Numerical ages after *Berggren et al.* [1985], except where amended by *Aubry et al.* [1988].

Thus we can subdivide the Cenozoic into essentially three broad benthic foraminiferal zones (Figure 2), as was earlier recognized by *Berggren et al.* [1992]. These zones can be recognized worldwide and over a very large depth range (upper to middle bathyal to upper abyssal), suggesting that the periods of change reflect major changes in the deep oceanic environment. The first of the zones ended at the rapid extinction at the end of the Paleocene. This extinction occurred at the beginning of a long-term warming trend of deep waters as seen in oxygen isotopic records of deep-sea benthic foraminifera (Figure 3), and at the end of a strong decrease of $\delta^{13}C$ values in surface and deep waters (Figure 4; events X and A of *Shackleton* [1986] and event C of *McGowran* [1990]). The more gradual periods of faunal change in the Eocene/Oligocene and early/middle Miocene overlapped with the two large, fast shifts to heavier values of $\delta^{18}O$ in deep-sea benthic foraminifera in the earliest Oligocene (35.8 Ma) and the middle Miocene (14.6 Ma), but the faunal change was more gradual, lasted longer, and started before the isotopic shifts.

THREE PERIODS OF BENTHIC FAUNAL CHANGE

Paleocene/Eocene Boundary Events

The Paleocene/Eocene boundary has not been generally recognized as a time of major biotic turnover, because generic extinction rates were low [e.g., *Raup and Sepkoski*, 1986]. These extinction patterns, however, largely show events in terrestrial and shallow marine environments and not in the deep ocean [*Thomas*, 1990*b*; *Kennett and Stott*, 1991]. The end of the Paleocene is within the lower reversed-polarity interval in Chron 24R [*Berggren et al.*, 1985; *Aubry et al.*, 1988], a period of profound changes in plate tectonic configuration [*Williams*, 1986; *McGowran*, 1989, 1989, 1991]. The northward motion of the Indian subcontinent slowed because of collision with Asia [*McGowran*,

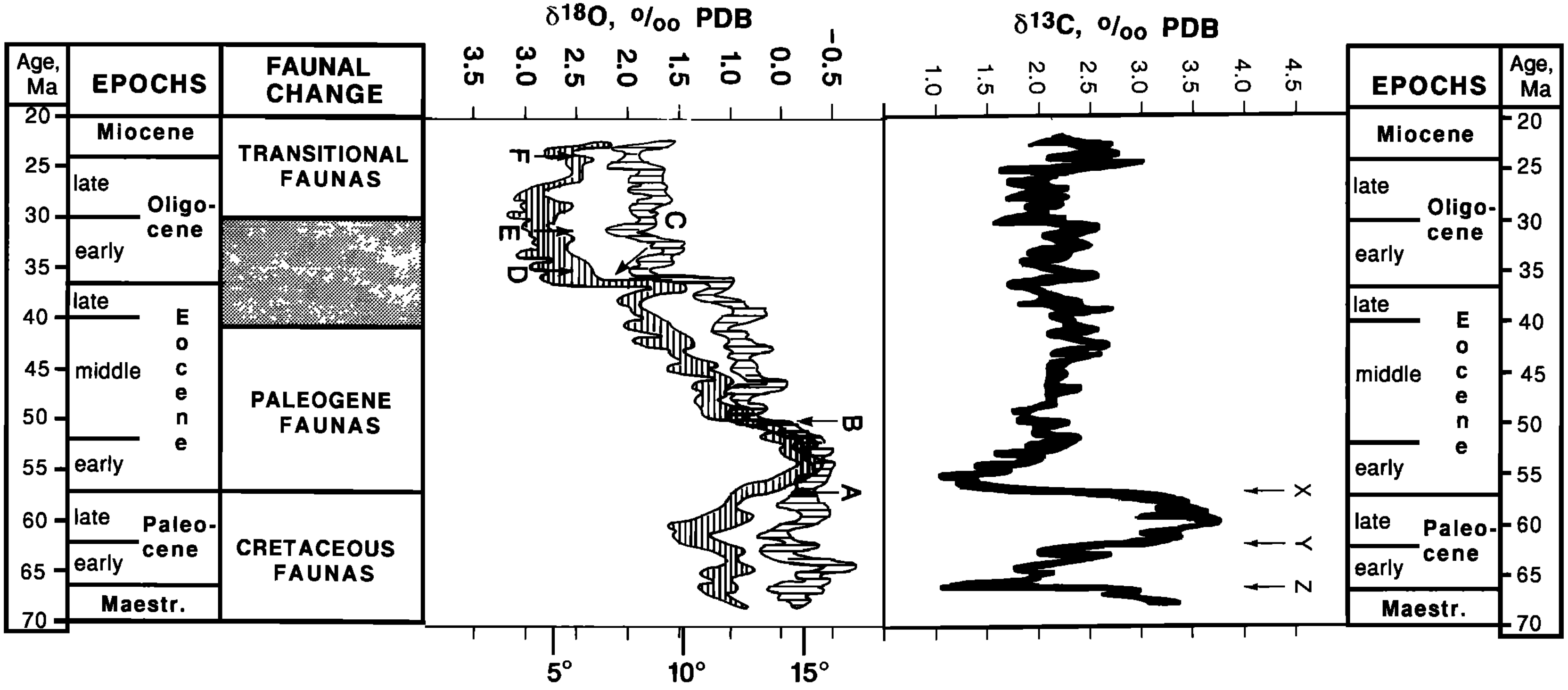

Fig. 4. Simplified benthic foraminiferal data for the Paleogene compared with oxygen isotopic data for bulk carbonate, reflecting surface water conditions (horizontal pattern) and benthic foraminifera (vertical pattern), and carbon isotopic data for bulk carbonate. Data and letters indicating events after *Shackleton* [1986]. Numerical ages after *Berggren et al.* [1985], except where amended by *Aubry et al.* [1989].

1991; *Klootwijk et al.*, 1991], the direction of subduction in the northern Pacific changed [*Goldfarb et al.*, 1991], and continental breakup started in the northern Atlantic. In this area, several millions of cubic kilometers of flood basalts erupted [*Roberts et al.*, 1984; *White and MacKenzie*, 1989; *White*, 1989; *Eldholm*, 1990] in unusually explosive eruptions [*Knox and Morton*, 1983, 1988; *Eldholm*, 1990]. The hydrothermal activity along the oceanic ridges in the Pacific was at its highest levels for the Cenozoic [*Leinen and Stakes*, 1979; *Owen and Rea*, 1985; *Leinen*, 1987; *Rea et al.*, 1990; *Kyte et al.*, 1992].

Rapid (10^3–10^4 years) environmental change occurred in the oceans, as observed in fluctuations in oxygen and carbon isotopic ratios of planktonic and benthic foraminifera. Carbon and oxygen isotopic values in benthic and planktonic foraminifera decreased by 1 to 2‰; the shift occurred within a few thousand years, and low values persisted for no longer than a few hundred thousand years [*Kennett and Stott*, 1991; *Pak and Miller*, 1992; *Lu and Keller*, 1992; E. Thomas and N. J. Shackleton, unpublished data]. The transient, large isotopic change has now been recognized in the Southern Ocean [*Kennett and Stott*, 1991; E. Thomas and N. J. Shackleton, unpublished data], the Atlantic Ocean [*Pak and Miller*, 1992; E. Thomas and N. J. Shackleton, unpublished data], the Atlantic Ocean [*Pak and Miller*, 1992; E. Thomas and N. J. Shackleton, unpublished data], the Indian Ocean [*Seto et al.*, 1991; *Barrera and Keller*, 1991; *Lu and Keller*, 1992; *Kennett*, 1991; *Thomas et al.*, 1992], and the equatorial Pacific [*Pak and Miller*, 1992] and is thus a global phenomenon. A coeval, large, transient shift to lower $\delta^{13}C$ values has also been observed in carbonate concretions and herbivore tooth enamel in North American land sections, suggesting that there was a major disturbance of the global carbon cycle, atmospheric as well as oceanic [*Koch et al.*, 1992]. The transient isotopic shift is at all oceanic sites coeval with the benthic foraminiferal extinction, suggesting that the extinction was coeval worldwide.

The short-term changes are superimposed on longer-term (10^6 years) changes [*Shackleton and Hall*, 1984, 1990; *Shackleton et al.*, 1984*a*, *b*, 1985; *Shackleton*, 1986, 1987; *Rea et al.*, 1990; *Thomas*, 1989, 1990*b*; *Corfield et al.*, 1991; *McGowran*, 1991; *Hovan and Rea*, 1992]; compare Figures 4–6. A reduction in the intensity of atmospheric circulation during this time was inferred from a sudden decrease of aeolian grain size, coeval with, or slightly before, the benthic foraminiferal extinction [*Rea et al.*, 1990; *Miller et at.*, 1987*b*; *Hovan and Rea*, 1992]. On land, there were major changes in mammalian faunas [*Butler et al.*, 1981, 1987; *Rea et al.*, 1990; *Koch et al.*, 1992]. Vegetation patterns suggest that the climate was warm and humid on the American continent [*Wolfe*, 1978; *Wing*, 1984; *Schmidt*, 1991; *Wing et al.*, 1991]. Increased abundance of the clay mineral kaolinite in deep-sea sediment at high southern latitudes suggests a more humid climate over the Antarctic continent [*Robert and Maillot*, 1990; *Robert and Chamley*, 1991; *Robert and Kennett*, 1992].

The benthic foraminiferal extinction at the end of the Paleocene was the only catastrophically sudden, global extinction of bathyal and upper abyssal benthic foraminifera during the last 75 m.y. [*Cushman*, 1946; *Braga et al.*, 1975; *Schnitker*, 1979*b*; *Tjalsma and Lohmann*, 1983; *Miller et al.*, 1987; *Berggren and Miller*, 1989; *Boltovskoy and Boltovskoy*, 1988, 1989; *Katz and Miller*, 1991; *Mackensen and Berggren*, 1992; *Nomura*, 1991; *Kaiho*, 1988, 1991; *Thomas*, 1989, 1990*a*, *b*; *Katz and Miller*, 1991]. There was no bathyal to abyssal benthic foraminiferal mass extinction at the Cretaceous/Tertiary boundary [*Beckmann*, 1960; *Dailey*, 1983; *Keller*, 1988; *Thomas*, 1990*b*; *Widmark and Malmgren*, 1992; *Kaiho*, 1992]. Paleocene deep-sea benthic foraminiferal faunas are so similar to Cretaceous faunas that early studies [e.g., *Cushman*, 1946] placed the Paleocene in the Cretaceous. The Paleocene $\delta^{13}C$ values of pelagic carbonates are more similar (heavy) to Cretaceous values than to values during the rest of the Cenozoic [*Shackleton*, 1987; *Corfield et al.*, 1991]. As to the development of the carbon cycle, the major break event in evolution toward the modern world seems to have occurred at the end of the Paleocene, and the major extinction at the end of the Cretaceous could be seen as a "freak accident," resulting from an impacting Apollo object [e.g., *Alvarez*, 1986]. The major environmental change in the latest Paleocene severely affected the deep oceans, which are volumetrically a very large part of the world's environment and of great importance to the heat balance of the ocean-atmosphere system.

At Maud Rise (Weddell Sea, Antarctica), Walvis Ridge (southern Atlantic Ocean), and Kerguelen Plateau (southern Indian Ocean), the extinction occurred slightly later than a decrease in $CaCO_3$ content of the sediments (Figure 6). At sites 525, 527 (Walvis Ridge), and 738 (Kerguelen), this decrease in $CaCO_3$ content from about 80–90% to 25–30% resulted in the presence of a dark brown clay layer across the extinction interval. The decrease in $CaCO_3$ content might reflect increased dissolution and a temporarily raised CCD. Directly after the extinction, both foraminiferal faunas and ostracode faunas are represented by small, thin-walled specimens, suggesting that the CCD did indeed move higher in the water column [*Thomas*, 1990*a*; P. Steineck, personal communication, 1991]. The decrease in $CaCO_3$ values, however, could also reflect a lowering of primary productivity, as is indicated to have occurred by a decrease in the $\delta^{13}C$ gradient between planktonic and benthic foraminifera [*Kennett and Stott*, 1991], and thus be similar in origin to the clay layers across the Cretaceous/Tertiary boundary.

A general, oceanwide drop in productivity was thought to have occurred in the latest Paleocene through

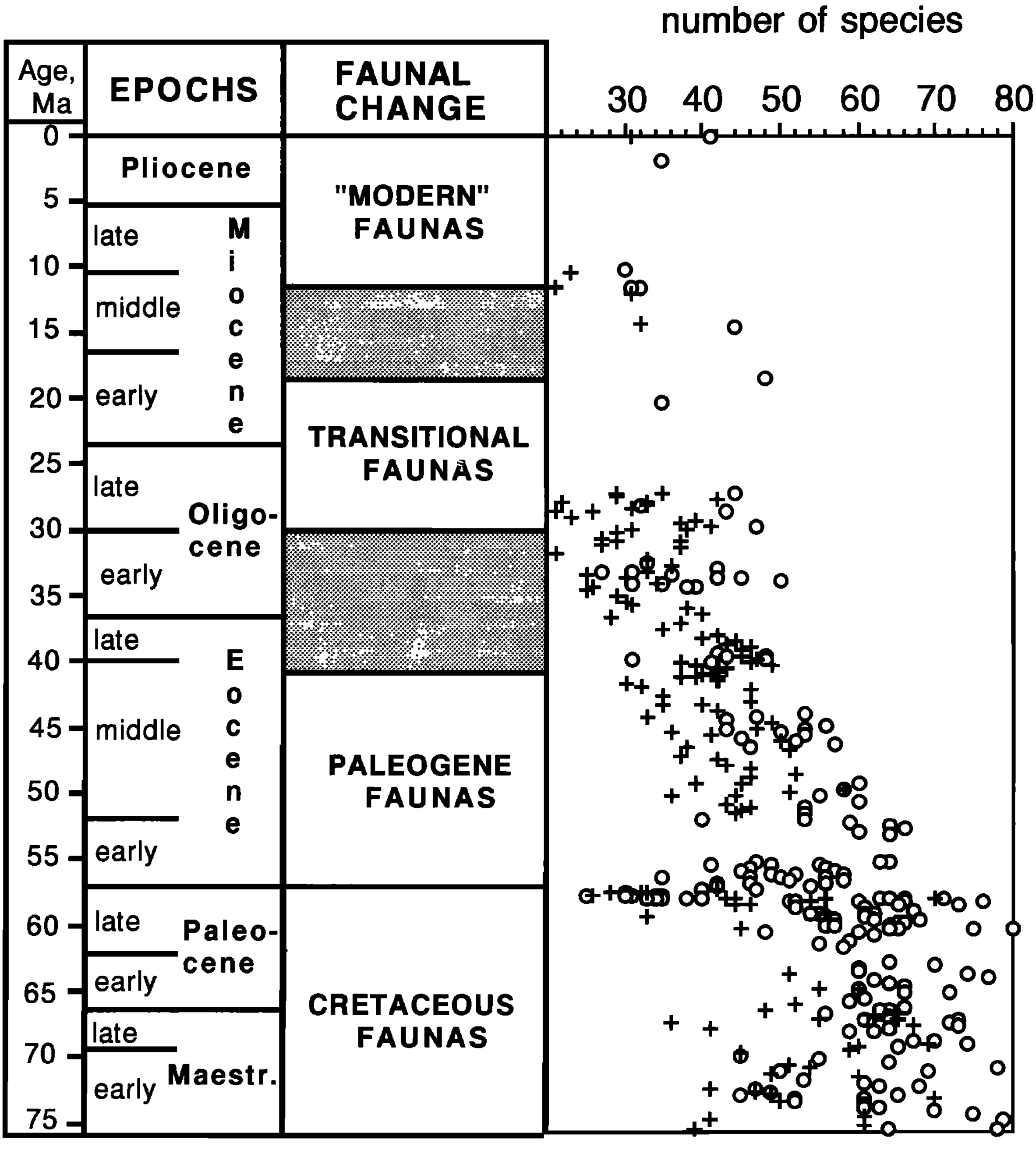

Fig. 5. Simplified benthic foraminiferal faunal changes compared with simple species richness at Site 689 (pluses) and Site 690 (circles) (Maud Rise, Antarctica). The number of species (normalized to 300 specimens) fluctuates strongly, but there is an overall decrease through the middle Eocene. Diversity is usually higher at the deeper site. Note the strong, sudden decrease in diversity at the end of the Paleocene.

earlier Eocene, from biogeographic patterns of planktonic foraminifera and carbon isotopes [*Shackleton et al.*, 1985; *Shackleton*, 1987; *Boersma and Premoli-Silva*, 1991; *Hallock et al.*, 1991]. Preliminary investigation of samples from Exmouth Plateau Site 762 and New Jersey margin Site 605, however, indicates that at these sites the productivity increased instead of decreased (E. Thomas, unpublished data), as inferred from absolute abundance of planktonic foraminifera, as well as from high relative abundance of chiloguembelinids.

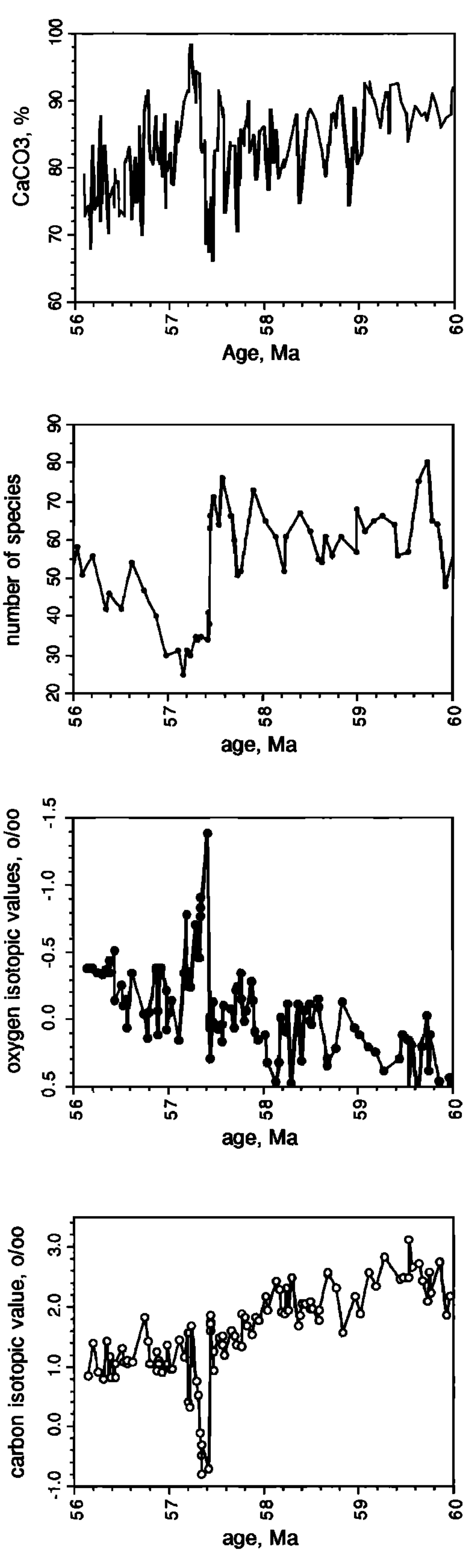

Fig. 6. Carbon and oxygen isotopic values of the benthic species *N. truempyi*, Site 690, compared with the number of species and percent $CaCO_3$ in the sediments. $CaCO_3$ values after *O'Connell* [1990]. Isotopic and faunal data after E. Thomas and N. J. Shackleton (work in preparation).

During the latest Paleocene deep-sea benthic foraminiferal extinction the diversity dropped sharply (by about 50%), and many long-lived, cosmopolitan and common species (such as *Gavelinella beccariiformis*) disappeared, as well as several typically Paleocene agglutinant species (e.g., *Tritaxia havanensis*, *Tritaxia paleocenica*, and *Dorothia oxycona*). After the extinction, low-diversity faunas were dominated by small, thin-walled specimens. Several species (e.g., *T. selmensis*) may have immigrated from shallower waters. Diversity never fully reached the levels of the Cretaceous and Paleogene, but it recovered in about 0.5 m.y. [*Berggren and Miller*, 1989; *Berggren et al.*, 1992; *Thomas*, 1989, 1990*a*, *b*, 1991; E. Thomas and N. J. Shackleton, work in preparation] (Figure 6). The faunal patterns after the extinction vary by site: at Atlantic sites the species *N. truempyi* is abundant after the extinction and is accompanied by small species such as *Abyssamina poagi*, *Quadrimorphina profunda*, and *Clinapertina planispira* [*Tjalsma and Lohmann*, 1983; *Miller et al.*, 1987*b*; *Berggren and Miller*, 1989; *Pak and Miller*, 1992; E. Thomas and N. J. Shackleton, work in preparation] (Figure 7). The faunal pattern at Site 762 in the eastern Indian Ocean resembles that at the Atlantic sites, but at Broken Ridge various *Anomalinoides* species dominated, together with *N. truempyi* [*Nomura*, 1991*a*]. At high-latitude sites on Maud Rise (Weddell Sea) and on the Kerguelen Plateau (Indian Ocean) the postextinction faunas are dominated by biserial species, most notably *Tappanina selmensis* [*Thomas*, 1989, 1990*a*, *b*, unpublished data]; at these sites *N. truempyi* disappeared temporarily, to return after about 150,000 years (Figure 7). The increase in the abundance of the biserial species at the high southern latitude sites could have resulted from lower oxygen content of the bottom waters and/or higher nutrient contents. It is difficult to see from where higher nutrient levels could have been derived at these sites, when there was a time of presumably low productivity (see above).

In land sections on Japan and New Zealand (deposited in the upper to middle bathyal zone) the typical Paleocene cosmopolitan taxa disappeared, and in the early Eocene more geographically limited faunas developed: the South Pacific–Atlantic–Tethyan Fauna (in New Zealand, Trinidad, and Italy) and the North Pacific Fauna (Japan and the Pacific coast of the United States [*Kaiho*, 1988].

The Paleocene deep oceans were populated by a cosmopolitan benthic fauna that occurred over a wide depth range, suggesting that a homogeneous water mass was present over large geographic and depth ranges [e.g., *Katz and Miller*, 1991]). After the extinction, more regional differences appeared, possibly indicating that deep and intermediate waters during the early to middle Eocene were more varied in physicochemical character and might have been derived from a number of source regions. This suggestion is in agreement with circulation

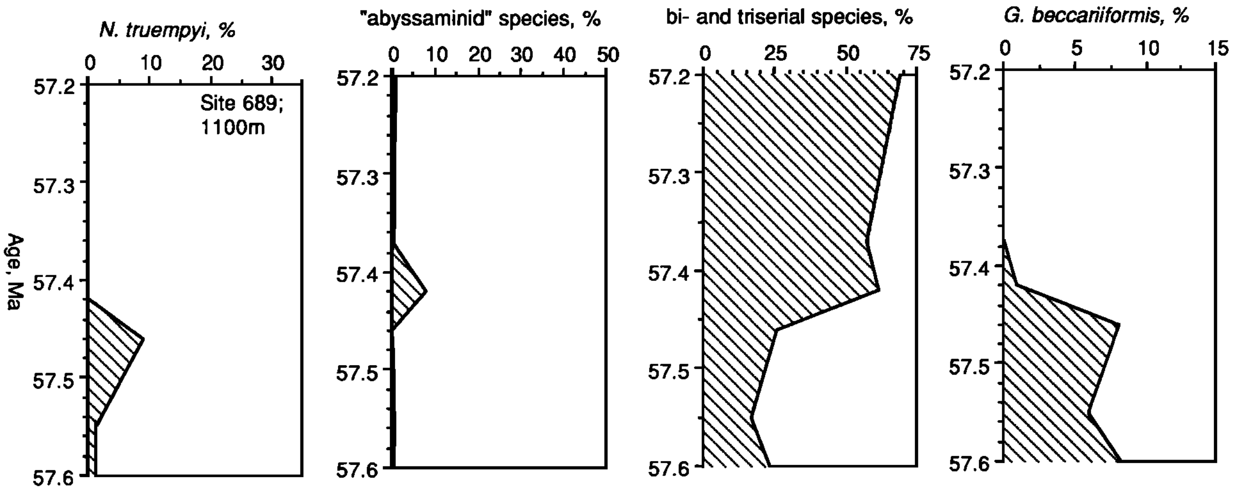

Fig. 7*a*

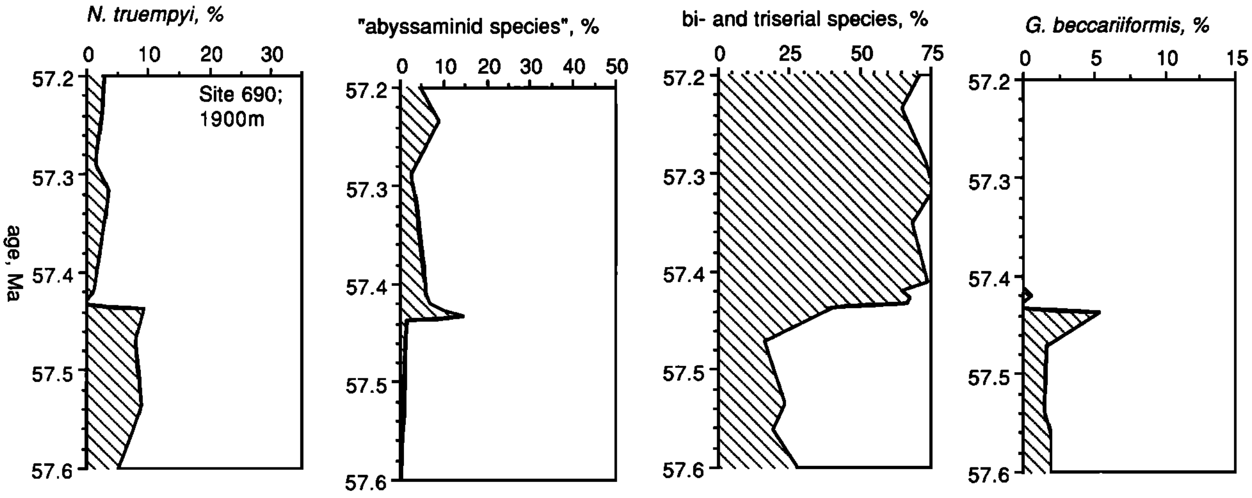

Fig. 7*b*

Fig. 7. Relative abundance of the most common groups of benthic foraminifera across the extinction at the end of the Paleocene, for two sites on Maud Rise (Weddell Sea: (*a*) Site 689, paleodepth of 1100 m; (*b*) Site 690, paleodepth of 1900 m) and two sites on Walvis Ridge (southern Atlantic Ocean: (*c*) Site 525, paleodepth of 1600 m; (*d*) Site 527, paleodepth of 3400 m).

models derived from carbon isotopic values [*Katz and Miller*, 1991; *Pak and Miller*, 1992; *Zachos et al.*, 1992*c*].

An explanation of the far-reaching, rapid climate change at the end of the Paleocene, and the benthic foraminiferal extinction, is not yet at hand [*McGowran*, 1991]. If the transient decrease in oxygen isotopic values at the time of extinction is explained wholly by increasing temperatures, there was a whole water column rise in temperature of 6°–7°C at Maud Rise [*Kennett and Stott*, 1990, 1991] (Figure 6). This increase was superimposed on a long-term increase in surface water temperatures at high latitudes, as indicated by the records of oxygen isotope values in bulk carbonate and planktonic foraminifera [*Stott et al.*, 1991] (Figure 6) and by the penetration of tropical species of planktonic foraminifera and nannofossils to high southern latitudes [*Haq et al.*, 1977; *Kennett*, 1978; *Stott and Kennett*, 1990; *Pospichal and Wise*, 1990; *Boersma and Premoli-Silva*, 1991; *Hallock et al.*, 1991; *Lu and Keller*, 1992; E. Thomas and N. J. Shackleton, work in preparation]. No mass extinctions were documented for planktonic foraminifera and calcareous nannoplankton, but both groups show global high diversity and peak turnover rates [*Haq et al.*, 1977; *Backman*, 1986*a*, *b*; *Corfield and Shackleton*, 1988; *Corfield and Cartlidge*, 1992; *Boersma and Premoli-Silva*, 1991; *Hallock et al.*, 1991; *Ottens and Nederbragt*, 1992; *Lu and Keller*, 1992].

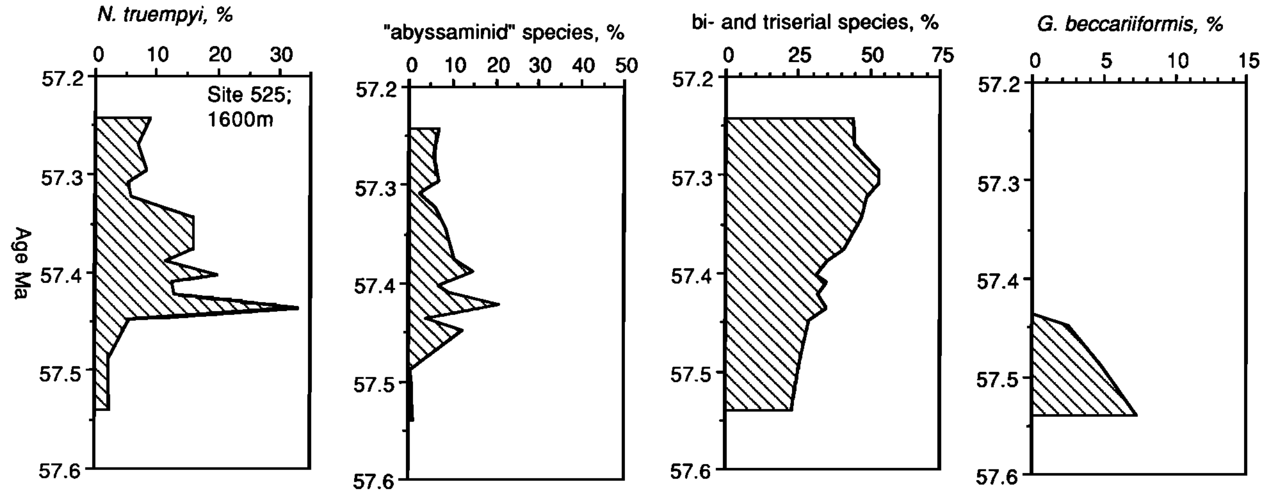

Fig. 7*c*

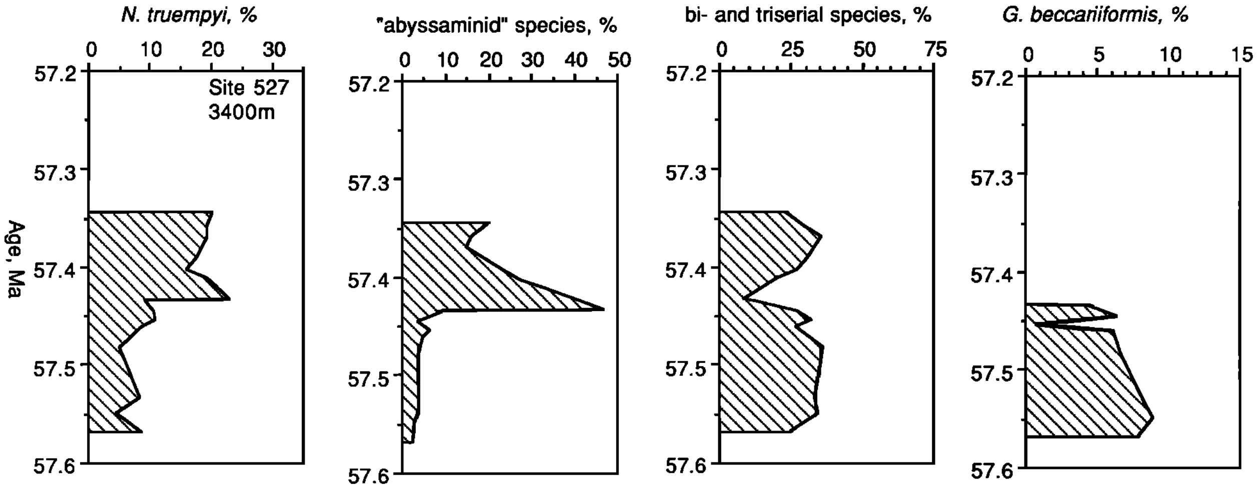

Fig. 7*d*

Calcareous nannoplankton shows extinction in *Fasciculithus* and *Rhomboaster* and radiations in *Discoaster* and *Cruciplacolithus* [*Romein*, 1979; *Backman*, 1986*a*].

Synchronous with the decrease in oxygen isotopic values there was a decrease of whole-ocean $\delta^{13}C$ values of dissolved carbonate by 1–2‰ (Figure 6; see also *Shackleton* [1987]). This change in carbon isotope value probably reflects at least partially an addition of isotopically light carbon to the oceans as well as the atmosphere (a so-called reservoir effect). It cannot purely be the result of productivity changes, because both surface and deep oceanic values are affected. The source of this isotopically light carbon was probably not continental weathering and erosion, because of the short time scale involved. It might have been the terrestrial biosphere, but there is no evidence for a collapse of terrestrial biota at the time [*Rea et al.*, 1990; *Wolfe*, 1978; *Wing*, 1984]. There is no indication of highly increased rate of erosion of shelf sediments, because there is no evidence for a precipitous drop in sea level at the time [*Haq et al.*, 1987].

Isotopically light carbon might possibly have been derived from CO_2 emissions related to the massive flood basaltic activity related to the opening of the North Atlantic and the very explosive phase of subaerial volcanism in the latest Paleocene [*Backman et al.*, 1984; *Roberts et al.*, 1984; *Berggren et al.*, 1985; *Aubry et al.*, 1986; *White and MacKenzie*, 1989; *Thomas*, 1991; *Thomas and Varekamp*, 1992]. Additional volcanic input might have been derived from the Pacific region, where there was a strong increase in hydrothermal activity at the end of the Paleocene [*Owen and Rea*, 1985; *Leinen*, 1987; *Leinen and Stakes*, 1979; *Lenotre et al.*, 1985; *Olivarez and Owen*, 1989; *Rea et al.*, 1990].

Speculating about the causes of the benthic foraminiferal extinction, we could hypothesize that a very strong pulse volcanic activity in the northern Atlantic caused high CO_2 levels in atmosphere and ocean [*Eldholm*, 1990], leading to transient global warming, especially at high latitudes [*Manabe and Bryan*, 1985]. The high temperatures at high latitudes could have led to formation of a low-density surface water layer, effectively preventing the formation of deep to intermediate waters at these latitudes. Deepwater circulation could then have changed, to formation dominantly at low latitudes. Such a change in circulation would have caused an overall decrease in oxygen content of the deep to intermediate waters. High relative abundances of biserial and triserial species of benthic foraminifera at high latitudes suggest that the oxygen content of the waters at depths from 1000 to 2000 m decreased at the time of extinction [*Kaiho*, 1988; *Thomas*, 1990*a*, *b*]. This change in oxygen content of deep to intermediate waters specifically at high latitudes could be the result of a combination of (1) lower oxygen contents of the deep to intermediate waters in the source area, as a result of higher temperatures, and (2) the longer route from the source area to high latitudes, and thus a longer "aging" time of the waters, resulting in even more increased levels of CO_2 and nutrients and decreased levels of O_2.

There is no indication in the benthic foraminiferal faunas that oxygen levels in the deep oceans, and especially at the high-latitude sites at Maud Rise in the Weddell Sea, were very low before the latest Paleocene extinction or after the early to earliest middle Eocene [*Thomas*, 1990*a*; *Kaiho*, 1988, 1991]. At Site 689 (1100-m paleodepth) and Site 690 (1900-m paleodepth) on Maud Rise (Antarctica) benthic faunas were very similar during the Eocene, suggesting that a similar water mass bathed both sites. This suggestion conflicts with oxygen isotopic evidence [*Kennett and Stott*, 1990]. At the deeper site, benthic oxygen isotopic values were lower during the Eocene, which has been interpreted as a temperature inversion due to the presence of warm, salty bottom water at Site 690. At present, this discrepancy cannot be explained, but it has been suggested that alternative explanations of the oxygen isotopic records are possible [*Wright and Miller*, 1992].

There is thus no benthic foraminiferal evidence that poorly oxygenated warm salty bottom water persisted in the Atlantic or Pacific oceans after the earliest middle Eocene. Carbon isotopic patterns suggest that the postulated source of warm salty bottom water was active for less than several 100,000 years, at least for waters in the Pacific and Atlantic oceans [*Miller et al.*, 1987*a*; *Katz and Miller*, 1991; *Pak and Miller*, 1992; *Zachos et al.*, 1992*c*].

Benthic faunal changes at the boundary between early and late Paleocene were at a much smaller scale than events at the end of the Paleocene, but they have been recognized at many sites, by many different researchers (Figure 2). These faunal changes were approximately coeval with the strong increase in carbon isotopic values of benthic and planktonic foraminifera and bulk carbonate (event Y of *Shackleton* [1987]; Figure 4). Presently, there is no clear explanation of this shift in carbon isotopes; it might at least result from increased productivity after recovery from the mass extinction at the end of the Cretaceous. The large isotopic shift, however, occurs in surface as well as deep waters [*Shackleton and Hall*, 1984; *Shackleton et al.*, 1984*a*, *b*, 1985; *Miller et al.*, 1987*b*; *Shackleton*, 1987; *Stott et al.*, 1990], and a reservoir effect must thus be involved. Possibly, the shift reflects the recovery of land biota after the end-Cretaceous extinction and thus storage of more light carbon in the terrestrial biosphere. More high-resolution records are needed over this interval that is commonly poorly recovered in DSDP and ODP drill holes.

The Middle Eocene Through Early Oligocene

The late Eocene was probably a period of cooling and growth of ice caps at high latitudes, as indicated by a relatively rapid increase in oxygen isotopic values of deep-sea benthic foraminifera [e.g., *Savin*, 1977; *Shackleton and Kennett*, 1975; *Kennett*, 1977; *Berger et al.*, 1981; *Corliss et al.*, 1984; *Shackleton*, 1984; *Corliss and Keigwin*, 1986; *Kennett and Stott*, 1990; *Zachos et al.*, 1992*c*]. It was also a period of high extinction rates in planktonic organisms as well as land faunas [*Raup and Sepkoski*, 1986; *Prothero and Berggren*, 1992]. Late Eocene extinctions have been suggested to have been caused by meteorite impact, because the tektites of the North American strewn field are upper Eocene [*Ganapathy*, 1982*a*, *b*; *Glass*, 1982; *Glass et al.*, 1983; *Glass and Zwart*, 1979; *Montanari*, 1990]. Overall, however, there is a poor correlation between the levels with tektites and those with extinctions. The tektites postdate the interval with most numerous extinctions in planktonic oceanic organisms and land fauna, the end of the middle Eocene, and predate the major shift in oxygen isotopic values of deep-sea benthic foraminifera [*Keller et al.*, 1983; *Keller*, 1983*a*, *b*, 1986; *MacLeod*, 1990; *Montanari*, 1990]. There is no short period of catastrophically sudden extinctions of deep-sea benthic foraminifera coeval with the short (100,000 years) oxygen isotopic shift in the earliest Oligocene [*Corliss*, 1981; *Thomas*, 1985, 1992; *Mackensen and Berggren*, 1992]. Benthic as well as planktonic foraminifera [*Keller*, 1983*a*, *b*; *Boersma and Premoli-Silva*, 1991; *Spezzaferri and Premoli-Silva*, 1991] show gradual extinction patterns from the middle middle Eocene on, without clusters of last appearances at the isotopic shift. The same pattern of gradual changes occurred in larger, neritic benthic foraminifera [*Hallock et al.*, 1991; *McGowran*, 1991] and smaller foraminifera in neritic sections [*McGowran*, 1987]. Extinctions occurred over a period of several millions of years from the middle Eocene into the early Oligocene in benthic foraminiferal faunas as well as in ostracode faunas [e.g., *Corliss*, 1981; *Tjalsma and Lohmann*, 1983; *Miller*, 1983; *Miller et al.*, 1984; *Boersma*, 1984, 1985, 1986; *Corliss and Keigwin*, 1986; *Kaiho*, 1988, 1991; *Berggren and Miller*, 1989; *Berggren et al.*, 1992; *Boltovskoy*, 1980; *Boltovskoy and Boltovskoy*, 1988, 1989; *Oberhänsli et al.*, 1991; *Thomas*, 1992] (see reviews in the work of *Douglas and Woodruff* [1981] and *Culver* [1987]).

Benthic foraminiferal faunal change started in the earliest part of the middle Eocene, between zones AB4 and AB5 [*Berggren and Miller*, 1989] and between assemblages 4B and 4A [*Thomas*, 1990*a*, 1992]; this change may not be worldwide and was certainly not noted by all investigators (Figure 2). It appears to be about coeval with oxygen isotopic event B of *Shackleton* [1986] (Figure 4), the first initiation of the gradual increase in $\delta^{18}O$ values in deep-sea benthic foraminiferal tests. Over this interval there are, however, few high-resolution data available. There is a parallel trend between the deep-sea benthic foraminiferal faunal events and the deep-sea oxygen isotopic record (Figure 5): the diversity of the faunas at Maud Rise sites 689 and 690, for instance, declines in parallel with the increase in oxygen isotopic values (compare Figure 5 with Figure 4).

At most sites the beginning of fast faunal overturn is placed somewhere near the end of the middle Eocene (Figure 2). At this time, there was a general decline in diversity at the high-latitude sites on Maud Rise (Antarctica; Figure 5). The cosmopolitan, very common species *Nuttallides truempyi* started to decline in abundance and migrated from bathyal to lower bathyal and abyssal depths [e.g., *Tjalsma and Lohmann*, 1983; *Miller*, 1983; *Miller et al.*, 1984; *Boersma*, 1984, 1985; *Corliss and Keigwin*, 1986; *Berggren and Miller*, 1989; *Mueller-Merz and Oberhänsli*, 1991; *Oberhänsli et al.*, 1991; *Thomas*, 1992]. The species had its last appearance at middle to lower bathyal depths at the end of the middle Eocene, at the end of the Eocene at lower bathyal to abyssal depths. A common phenomenon at many sites is the decline in relative abundance (to $<10\%$) of buliminid species at lower bathyal to abyssal depths, especially of the larger, heavily calcified species [*Miller*, 1983; *Miller et al.*, 1984; *Boersma*, 1984, 1985; *Thomas*, 1989, 1990*a*, 1992; *Mueller-Merz and Oberhänsli*, 1991; *Oberhänsli et al.*, 1991].

There does not appear to be a time correlation between these benthic faunal changes and fluctuation in large-scale features of the carbon isotopic record, but *Diester-Haass* [1991] and *Thomas* [1992] argued that there is evidence for increased surface productivity at the timing of the oxygen isotopic shift, as well as at the beginning of the benthic foraminiferal faunal change, at Site 689 (Maud Rise, Antarctica). *Zachos et al.* [1992*c*] recognized a change in carbon isotopes in Indian Ocean sites, which suggests increased productivity, starting shortly before the oxygen isotopic shift. *Thomas* [1992] correlated the gradual faunal change at Maud Rise over this period primarily with the gradual decrease in temperatures of the deep waters and the concomitant increase in corrosivity of the waters as a result of increased solubility of CO_2. This increase in corrosivity is obvious at high-latitude sites, where the CCD decreased precipitously [*Barker et al.*, 1988], but at many sites in the Pacific and Atlantic oceans the CCD increased in the earliest Oligocene [*van Andel*, 1975]. Possibly the increased ventilation of the deep ocean, which started at the end of the Eocene [*Miller et al.*, 1987*a*], resulted in decreased levels of CO_2 in deep waters at middle to low latitudes and thus counteracted the increase in CO_2 levels resulting from a temperature drop at high latitudes. The decrease in the relative abundance of the *Bulimina* species could then be, at least partly, caused by increasing levels of oxygenation of the deep waters. This increased oxygenation could have resulted in more thorough oxidation of organic material, leaving less nutrients available for the benthic foraminiferal faunas.

The Early to Middle Miocene

Major changes occurred in the biosphere and the Earth's climate in the middle Miocene. The oxygen isotopic records indicate that some combination of rapid (100,000 years) ice growth and temperature decline at high latitudes occurred, as at the end of the Eocene [e.g., *Shackleton and Kennett*, 1975; *Kennett and Shackleton*, 1976; *Kennett*, 1977; *Savin*, 1977; *Berger et al.*, 1981; *Douglas and Woodruff*, 1981; *Woodruff and Douglas*, 1981; *Kemp*, 1983; *Savin et al.*, 1985; *Shackleton*, 1984; *Berger and Vincent*, 1985; *Vincent and Killingley*, 1985; *Webb*, 1990]. High-resolution studies have shown that during the Miocene isotopic changes, short-term high $\delta^{13}C$ values appear to occur at times of high $\delta^{18}O$ values, suggesting that periods of cooling and ice buildup were associated with rapid burial of organic material and lowered atmospheric CO_2 levels [*Berger and Vincent*, 1985; *Woodruff and Savin*, 1991]. The middle Miocene has been named as a period of faunal overturn in studies that argue for periodicity in extinction patterns [e.g., *Raup and Sepkoski*, 1986]. At some locations, iridium enrichments have been detected at middle Miocene levels, but no global pattern has been found [*Asaro et al.*, 1988].

Benthic faunal change clearly occurred during this time (about 17–13 Ma) [*Berggren and Miller*, 1989]) and started before the oxygen isotopic increase [*Thomas*, 1985, 1986; *Woodruff*, 1985; *Thomas and Vincent*, 1987, 1988; *Miller and Katz*, 1987; *Boltovskoy and Boltovskoy*, 1988, 1989; *Nomura*, 1991*b*; *Nomura et al.*, 1992]. The faunal change affected about 20% of the species [*Boltovskoy and Boltovskoy*, 1988; *Thomas*, 1986*b*]. The widespread species *Cibicidoides wuellerstorfi* and *Pyrgo murrhyna* evolved toward the end of the period of faunal change [e.g., *Thomas*, 1985; *Bolt-*

ovskoy, 1980, 1987; *Boltovskoy and Boltovskoy*, 1988; *Thomas and Vincent*, 1987, 1988; *Woodruff and Savin*, 1991]. Earlier in the period, uniserial species such as nodosariids and pleurostomellids, that had been numerous during the Cretaceous and the Paleogene, decreased in relative abundance, and miliolid species increased in relative abundance in the deep sea [*Boltovskoy and Boltovskoy*, 1988; *Thomas*, 1986]. The environmental significance of these faunal data is not clear.

The offset in timing between oxygen isotopic and benthic faunal change with the benthic faunal change leading cannot result from problems in correlation, because it has been noted at many sites where faunal and isotopic data were obtained from the same cores or even the same samples [*Woodruff*, 1985; *Miller and Katz*, 1987; *Thomas and Vincent*, 1987]. There might be a correlation in time between the initiation of benthic foraminiferal faunal change (about 17 Ma) and the early Miocene (Chron 16) carbon shift (the so-called Monterey event of *Berger and Vincent* [1985, 1986] *Miller and Fairbanks*, 1985; *Thomas and Vincent*, 1987]). The evolution of *C. wuellerstorfi* (15.0–15.6 Ma [*Thomas*, 1985] and 15.3 Ma [*Woodruff and Savin*, 1991]) occurred in the equatorial Pacific, from where the species spread out into the world's oceans fairly late in the period of faunal overturn.

It has been suggested that this faunal overturn was caused by changes in surface-ocean productivity or flux of organic carbon to the ocean floor [*Miller and Katz*, 1987; *Thomas and Vincent*, 1987; *Boltovskoy and Boltovskoy*, 1987]. *Thomas and Vincent* [1987] suggested that faunal overturn was influenced by changes in productivity as well as changes in corrosivity of the deep waters. *Woodruff and Savin* [1989], however, suggested that changing patterns of deepwater circulation, especially the volume of the outflow of salty, high-density waters at intermediate depths from the closing eastern end of the Tethys Ocean, might have been involved. Deep-sea benthic foraminiferal assemblages at DSDP sites 608 and 610 in the northeastern Atlantic suggest that just before the carbon isotopic shift there were episodes of at least local, sluggish circulation leading to poorly oxygenated basins in the North Atlantic [*Thomas*, 1986*b*]. For a period of about 1 m.y. (19–18 Ma), benthic faunas at these sites were strongly dominated by small, thin-walled bolivinids. Recently, these episodes of bolivinid-dominated faunas have also been recognized at other sites in the northern, southern, and equatorial Atlantic [*Smart*, 1991] but not in the equatorial Pacific [*Thomas*, 1985; *Woodruff*, 1985].

Presently, there is thus no clear, unequivocal correlation between faunal and oxygen or carbon isotopic events in the early to middle Miocene, but there appears to be some correlation between faunal change and changes in productivity.

DISCUSSION AND SPECULATION

In theory, we might expect global deep-sea benthic foraminiferal faunal change to occur as a result of change in source area of the deepwater masses and change in character of the waters in the source areas. This signal is expected to have become complicated by changes in productivity, which in turn might also have been influenced by changing oceanic circulation patterns.

Rapid Faunal Change

Commonly, benthic faunal change occurs rapidly at one site, as one assemblage is replaced by another. This type of faunal change is reversible, and several assemblages may alternate at one site in time [e.g., *Mueller-Merz and Oberhänsli*, 1991; *Nomura*, 1991*b*; *Oberhänsli et al.*, 1991]. There may appear to be a general correlation in the timing of these faunal changes from one site to another, but in the presence of high-resolution biostratigraphic and/or magnetostratigraphic data the timing proves to differ from site to site. In the absence of local effects such as tectonic changes in water depth at the site, we can interpret these changes most probably as indicating that the depth of boundaries between water masses at the site(s) fluctuated over time [e.g., *Oberhänsli et al.*, 1991], owing to changes in volume of the different water masses. Such an explanation has been proposed for the glacial/interglacial benthic foraminiferal faunal change in the northern Atlantic Ocean [*Streeter*, 1973; *Streeter and Shackleton*, 1979; *Schnitker*, 1974, 1979*a*].

To illustrate the pattern of faunal change expected from such a change in circulation, compare Figures 8*a* and 8*b*. The volume of deep water from the northern high latitude is greater in Figure 8*a* than in Figure 8*b*. If Figure 8*b* type circulation changes to Figure 8*a* type circulation, faunal change may be expected at site C, but not at other sites. The faunal changes will be rapid, because the water mass boundary passes quickly over each location. We need data on (at least) several sites at different depths, in the same general area, to decipher such motion of water mass boundaries [*Woodruff*, 1985; *Thomas*, 1986*a*; *Kurihara and Kennett*, 1988]. The faunal changes are diachronous from site to site, because the timing of the passage of water mass boundaries over different sites differs by site.

Gradual Faunal Change

Another type of faunal change occurs over millions of years and can be recognized globally. During the Cenozoic there were at least two times of such faunal change: (1) the late Eocene through early Oligocene and (2) the early middle Miocene through late middle Miocene [e.g., *Corliss*, 1981; *Tjalsma and Lohmann*, 1983; *Miller*, 1983; *Miller et al.*, 1984; *Boersma*, 1984, 1985;

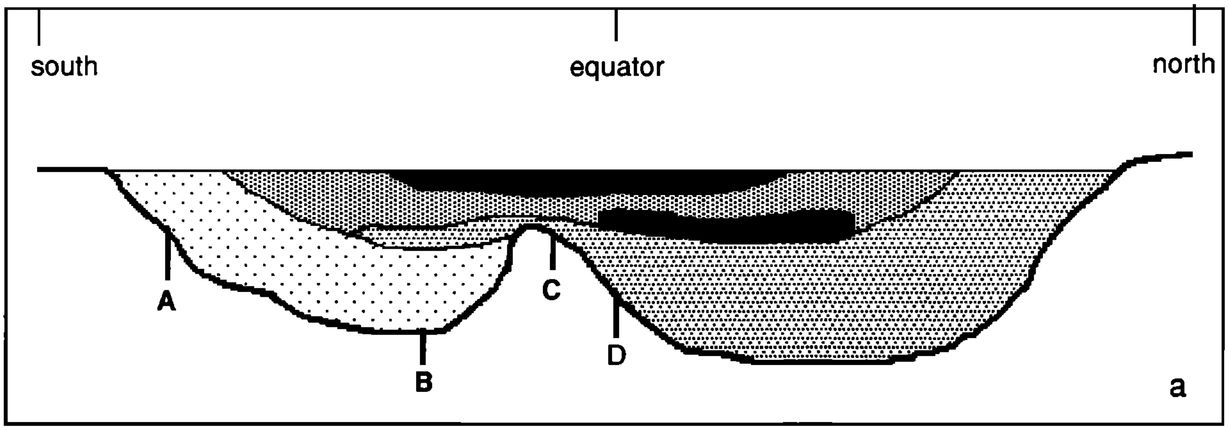

Fig. 8. North-south cross sections across an imaginary ocean to illustrate some of the many possible patterns of deepwater circulation. Letters A through D show possible locations from which deep-sea benthic foraminifera are studied and how these are affected by the changes in deepwater circulation.

Fig. 8*a*. Deep water formed dominantly at high northern as well as southern latitudes, with waters from the northern source filling the larger part of the ocean; the volume of warm salty deep water is very small (similar to the present Atlantic Ocean).

Corliss and Keigwin, 1986; *Berggren and Miller*, 1989; *Boltovskoy*, 1980; *Boltovskoy and Boltovskoy*, 1988, 1989; *Oberhänsli et al.*, 1991; *Douglas and Woodruff*, 1981; *Culver*, 1987; *Woodruff and Douglas*, 1981; *Thomas*, 1985, 1986*a*, *b*, 1992; *Woodruff*, 1985; *Murray et al.*, 1986; *Schnitker*, 1986; *Thomas and Vincent*, 1987, 1988; *Miller and Katz*, 1987]. Each of these periods of faunal overturn has been correlated, at least tentatively, with changes in productivity. These gradual changes might be influenced by gradual climate change in the areas of deep or intermediate water formation. During gradual climate change in the source region of the deep water, the surface waters change, and thus the deep waters that form from these surface waters (e.g., temperature, O_2 content, CO_2 content, and preformed nutrient content) change.

Oxygen isotopic records suggest that deep waters cooled gradually from the middle Eocene on [e.g., *Miller et al.*, 1987*a*], while benthic faunas suffered gradual turnover at the same time (Figure 5). The benthic faunal overturn can thus be thought to reflect the gradual environmental change. This speculation suggests that at least part of the earliest Oligocene oxygen isotopic shifts must have been related to ice volume increase, because there was no rapid benthic extinction coeval with the isotopic shift. Benthic fora-

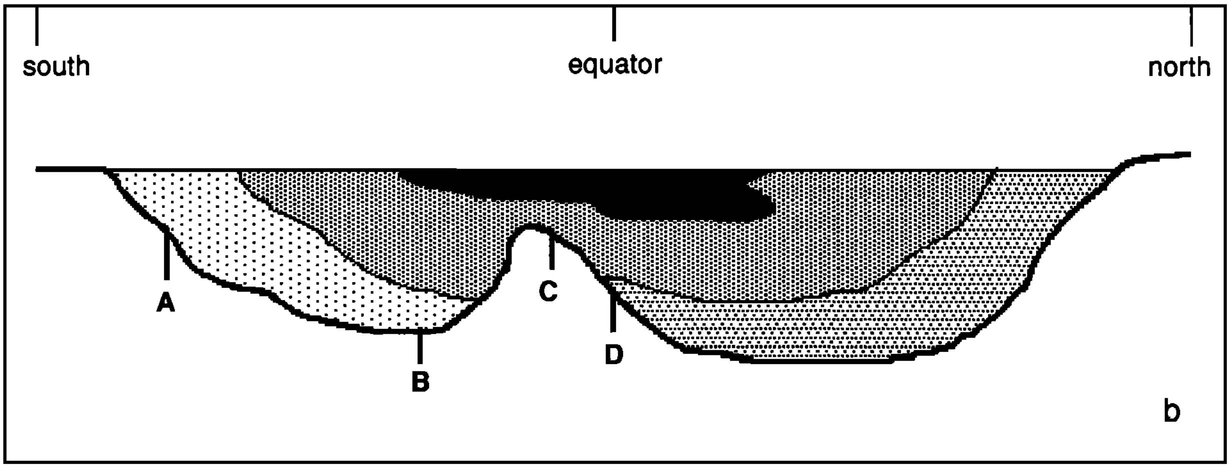

Fig. 8*b*. Deep water formed at high northern as well as southern latitudes, but less dominated by the northern source (possibly similar to the Atlantic during the last glacial). Note that after a change in circulation from Figure 8*a* to Figure 8*b* the faunas at site C are affected, but not those at sites A, B, and D.

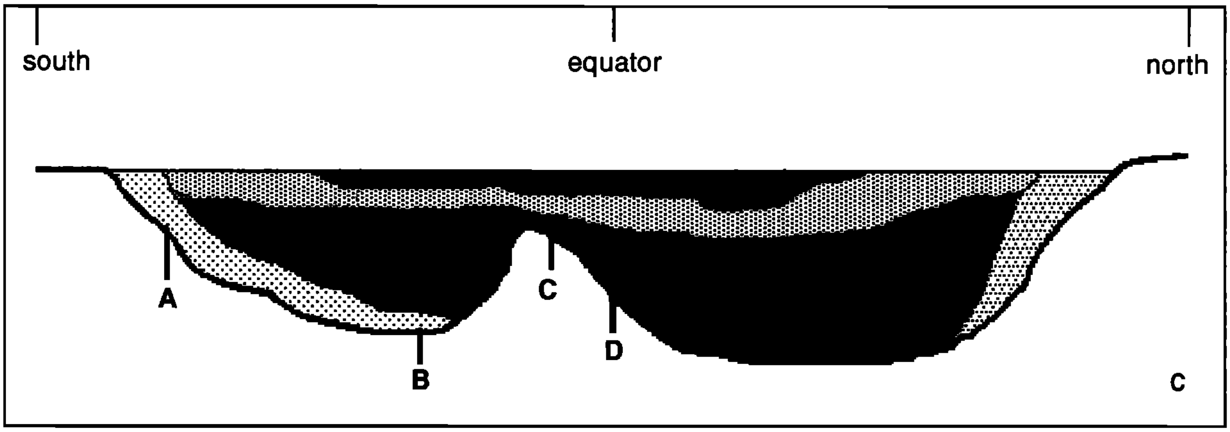

Fig. 8*c*. A hypothetical ocean in which the role of warm salty bottom water is much larger than that in Figures 8*a* and 8*b*, but deepest waters are formed at high latitudes. We can envisage this situation as well as the possibility that the high-latitude waters flow over the warm salty bottom waters (the Proteus Ocean [*Kennett and Stott*, 1990]). A change from Figure 8*c* to Figure 8*a* or 8*b* will be reflected in faunas at sites C and D, but not at sites A and B.

miniferal faunas would "see" the gradual decrease in temperatures leading up to the rapid isotopic shift, but not the increased ice volume on land. The correlation of benthic faunal change and decreasing high-latitude temperatures might be partially indirect and reflect changes in oceanic productivity resulting from high-latitude cooling and changes in the rate of oceanic turnover.

Global Mass Extinctions

This type of faunal change has been recorded only in the latest Paleocene, not at any other time in the Maestrichtian through Recent [*Schnitker*, 1979*b*; *Tjalsma and Lohmann*, 1983; *Miller et al.*, 1987; *Kaiho*, 1988; *Thomas*, 1989, 1990*a*, *b*; Nomura, 1991*a*]. The extinction is rapid (10^2–10^3 years) and probably globally synchronous. An explanation for such an oceanwide event is hard to find. We need to speculate that there were either very fast changes in physicochemical character of the surface water in all deepwater source regions or a change in source area of deep waters affecting the circulation patterns of the whole ocean, not just one oceanic basin. After all, the Pleistocene glacial-interglacial circulation changes in the North Atlantic did not result in a mass extinction of deep-sea benthic foraminifera; assemblages appear to have migrated with the water masses [*Streeter*, 1974; *Streeter and Shackleton*, 1979; *Schnitker*, 1974]. The rarity of global extinction suggests that complete turnovers in deepwater

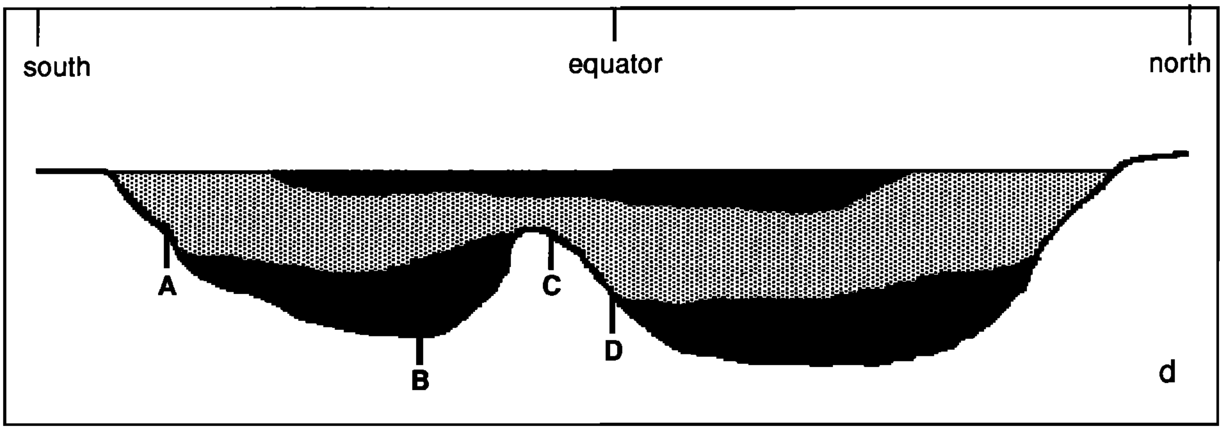

Fig. 8*d*. A hypothetical ocean in which the deepest parts of the basins are filled with warm salty deep water, but the largest parts of the oceans (including almost the entire depth range above the CCD) are filled with intermediate waters derived from high latitudes. In such an ocean the faunas over wide areas and depth ranges would be very similar (sites A, C, and D are in the same water mass), except for the deep regions of the basins (site B) from which we do not have information.

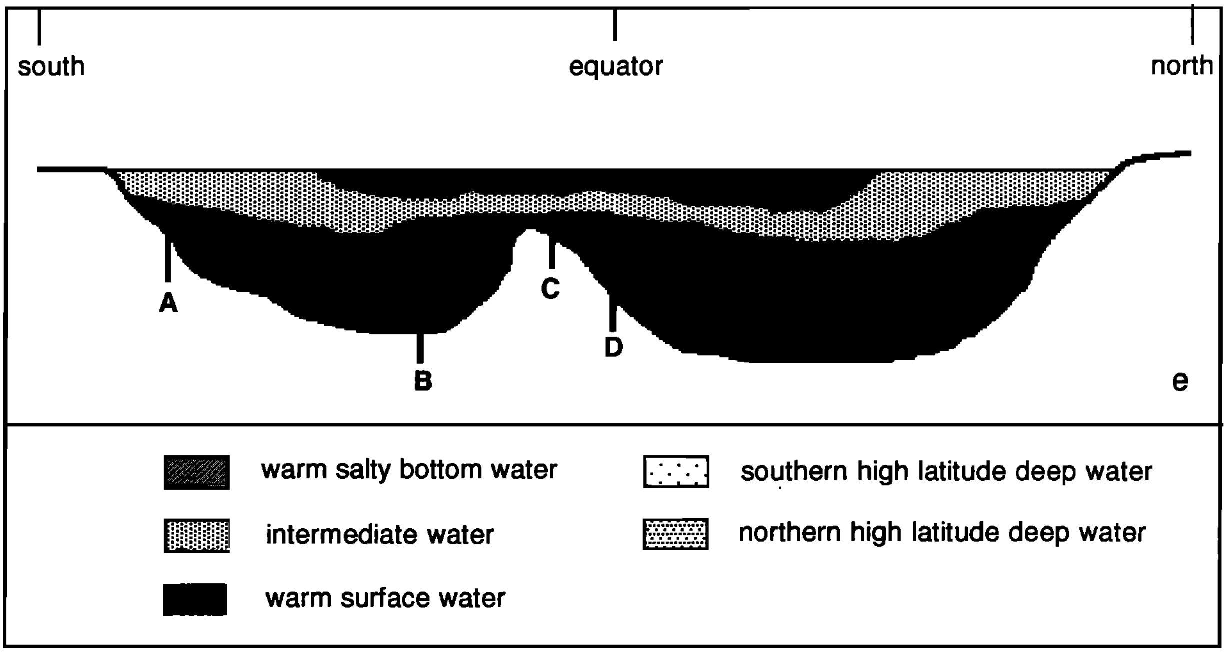

Fig. 8*e*. Similar to the proposed ocean in Figure 8*d*, but the overall volume of the warm salty bottom water is much larger, and the intermediate waters are limited to a thin zone. Note that a change from the situation in Figure 8*e* to that in Figure 8*d* (and the reverse) will result in faunal change at sites A, C, and D from which we can obtain information.

circulation during Late Cretaceous through Cenozoic occurred only in the latest Paleocene.

We can envisage several possible configurations of "reversed" circulation in the deep oceans, as indicated in Figures 8*c* and 8*d*. A change from the type of oceanic circulation as shown in Figures 8*a* and 8*b* to one as shown in Figures 8*d* and 8*e* could occur, possibly through an intermediary stage similar to that in Figure 8*c* [see also *Kennett and Stott*, 1990]. Such large-scale circulation changes might affect faunas at all locations (hypothetical sites A through D). Faunal change might be exacerbated because of changes in locations of upwelling and in nutrient content of upwelled water and thus of the location of areas of high surface productivity.

Such a reversal in circulation of the deep to intermediate waters might have been triggered by the warming of surface waters at high latitudes (as indicated by oxygen isotopic as well as faunal data), resulting in high-latitude surface waters with a low density. This could have been aggravated by an increasingly humid climate and more precipitation at high latitudes, as indicated by changes in clay mineral assemblages at Maud Rise [*Robert and Maillot*, 1990; *Robert and Chamley*, 1991; *Robert and Kennett*, 1992]. Such a reversed circulation could have influenced upwelling patterns and nutrient availability to the plankton and thus plankton evolutionary patterns.

CONCLUSIONS

Benthic foraminiferal faunal data do not support the hypothesis that "warm salty bottom water" existed over large parts of the oceans during most of the Late Cretaceous and the Paleogene. They do support the possibility that such waters were dominant in the Atlantic and Pacific during one or a few such episodes during the early to earliest middle Eocene. The initiation of such an event of warm salty deepwater dominance in the oceans might have caused the extinction in the latest Paleocene. Benthic foraminiferal faunal data suggest that at least parts of the oxygen isotopic shifts to heavier values in the earliest Oligocene and the middle Miocene represent the buildup of ice volume.

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REFERENCES

Adams, C. G., D. E. Lee, and B. R. Rosen, Conflicting isotopic and biotic evidence for tropical sea-surface temperature during the Tertiary, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *77*, 289–313, 1990.

Alvarez, W., Toward a theory of impact crises, *Eos Trans. AGU*, *67*, 649–658, 1986.

Asaro, F., W. Alvarez, H. V. Michel, L. W. Alvarez, M. H. Anders, A. Montanari, and J. P. Kennett, Possible worldwide middle Miocene iridium anomaly and its relationship to periodicity of impacts and extinctions (abstract), in *Global Catastrophes in Earth History: Conference on Impacts, Volcanism and Mass Mortality, Abstracts Volume*, pp. 6–7, Lunar and Planetary Institute, Houston, Tex., 1988.

Aubry, M.-P., E. A. Hailwood, and H. A. Townsend, Magnetic and calcareous nannofossil stratigraphy of lower Paleogene formations of the Hampshire and London basins, *J. Geol. Soc. London*, *143*, 729–735, 1986.

Aubry, M.-P., W. A. Berggren, D. V. Kent, J. J. Flynn, K. D. Klitgord, J. D. Obradovich, and D. R. Prothero, Paleogene chronology: An integrated approach, *Paleoceanography*, *3*, 707–742, 1988.

Backman, J., Late Paleocene to middle Eocene calcareous nannofossil biochronology from the Shatsky Rise, Walvis Ridge and Italy, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *57*, 43–59, 1986*a*.

Backman, J., Accumulation patterns of Tertiary calcareous nannofossils around extinctions, *Geol. Rundsch.*, *75*, 185–196, 1986*b*.

Backman, J., A. C. Morton, D. G. Roberts, S. Brown, K. Krumsieck, and R. M. MacIntyre, Geochronology of the lower Eocene and upper Paleocene sequences of Leg 81, *Initial Rep. Deep Sea Drill. Proj.*, *81*, 877–882, 1984.

Barker, P. F., et al., Leg 113, *Proc. Ocean Drill. Program Initial Rep.*, *113*, 785 pp., 1988.

Barrera, E., and G. Keller, Late Paleocene to early Eocene climatic and oceanographic events in the Antarctic Indian Ocean (abstract), *Geol. Soc. Am. Abstr. Programs*, *23*(5), A179, 1991.

Barrera, E., B. Huber, S. M. Savin, and P. N. Webb, Antarctic marine temperatures: Late Campanian through early Paleocene, *Paleoceanography*, *2*, 21–48, 1987.

Barron, E. J., Explanations of the Tertiary global cooling trend, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *50*, 45–61, 1985.

Barron, E. J., Eocene equator-to-pole surface ocean temperatures: A significant climate problem?, *Paleoceanography*, *2*, 729–740, 1987.

Barron, E. J., and W. H. Peterson, The Cenozoic ocean circulation based on ocean general circulation model results, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 1–18, 1991.

Barron, J. A., and J. G. Baldauf, Tertiary cooling steps and paleoproductivity as reflected by diatoms and biosiliceous sediments, in *Productivity of the Oceans: Past and Present*, edited by W. H. Berger, V. S. Smetacek, and G. Wefer, pp. 341–345, John Wiley, New York, 1989.

Barron, J. A., B. L. Larsen, and J. Baldauf, Evidence for late Eocene–early Oligocene Antarctic glaciation and observations on late Neogene glacial history of Antarctica: Results from ODP Leg 119, *Proc. Ocean Drill. Program Sci. Results*, *119*, 869–891, 1991.

Beckmann, J.-P., Distribution of benthonic foraminifera at the Cretaceous-Tertiary boundary of Trinidad (West Indies), in *Report of the 21st Session, Norden, Part 5: The Cretaceous-Tertiary Boundary*, pp. 57–69, International Geological Congress, Norden, Denmark, 1960.

Benson, R. H., The origin of the psychrosphere as recorded in changes of deep-sea ostracode assemblages, *Lethaia*, *8*, 69–83, 1975.

Berger, W. H., Paleoceanography: The deep sea record, in *The Sea*, vol. 7, *The Oceanic Lithosphere*, edited by C. Emíliani, pp. 1437–1519, John Wiley, New York, 1981.

Berger, W. H., and E. Vincent, Carbon dioxide and polar cooling in the Miocene: The Monterey hypothesis, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present, Geophys. Monogr. Ser.*, vol. 32, edited by E. T. Sundquist and W. S. Broecker, pp. 455–468, AGU, Washington, D. C., 1985.

Berger, W. H., and E. Vincent, Deep-sea carbonates: Reading the carbon isotope signal, *Geol. Rundsch.*, *75*, 249–269, 1986.

Berger, W. H., E. Vincent, and H. Thierstein, The deep-sea record: Major steps in Cenozoic ocean evolution, *Spec. Publ. Soc. Econ. Paleontol. Mineral.*, *32*, 489–504, 1981.

Berggren, W. A., and K. G. Miller, Cenozoic bathyal and abyssal calcareous benthic foraminiferal zonation, *Micropaleontology*, *35*, 308–320, 1989.

Berggren, W. A., D. V. Kent, J. J. Flynn, and J. A. Van Couvering, Cenozoic geochronology, *Geol. Soc. Am. Bull.*, *96*, 1407–1418, 1985.

Berggren, W. A., M. E. Katz, and K. G. Miller, Cenozoic deep-sea benthic foraminifera: A tale of three turnovers, in *Fourth International Symposium on Benthic Foraminifera*, edited by T. Saito, Yamagata University, Yamagata, Japan, in press, 1992.

Bernard, J. M., Characteristic assemblages and morphologies of benthic foraminifera, from anoxic, organic-rich deposits: Jurassic through Holocene, *J. Foraminiferal Res.*, *16*, 207–215, 1986.

Boersma, A., Oligocene and other Tertiary benthic foraminifera from a depth traverse down Walvis Ridge, Deep Sea Drilling Project Leg 74, *Initial Rep. Deep Sea Drill. Proj.*, *75*, 1273–1300, 1984.

Boersma, A., Oligocene benthic foraminifers from North Atlantic sites: Benthic foraminifers as water-mass indexes in the North and South Atlantic, *Initial Rep. Deep Sea Drill. Proj.*, *82*, 611–628, 1985.

Boersma, A., Eocene-Oligocene Atlantic paleo-oceanography, using benthic foraminifera, in *Terminal Eocene Events*, edited by C. Pomerol and I. Premoli-Silva, pp. 225–236, Elsevier, New York, 1986.

Boersma, A., and I. Premoli-Silva, Distribution of Paleogene planktonic foraminifera—Analogies with the Recent?, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 29–48, 1991.

Boersma, A., I. Premoli-Silva, and N. J. Shackleton, Atlantic Eocene planktonic foraminiferal paleohydrographic indicators and stable isotope stratigraphy, *Paleoceanography*, *2*, 287–331, 1987.

Boltovskoy, E., On the benthonic bathyal-abyssal foraminifera as stratigraphic guide to fossils, *J. Foraminiferal Res.*, *10*, 163–172, 1980.

Boltovskoy, E., Tertiary benthic foraminifera in bathyal deposits of the Quaternary world ocean, *J. Foraminiferal Res.*, *17*, 279–285, 1987.

Boltovskoy, E., and D. Boltovskoy, Cenozoic deep-sea benthic foraminifera: Faunal turnovers and paleobiographic differences, *Rev. Micropaleontol.*, *31*, 67–84, 1988.

Boltovskoy, E., and D. Boltovskoy, Paleocene-Pleistocene benthic foraminiferal evidence of major paleoceanographic events in the eastern South Atlantic (DSDP Site 525, Walvis Ridge), *Mar. Micropaleontol.*, *14*, 283–316, 1989.

Braga, G., R. deBiase, A. Grunig, and F. Proto-Decima, Foraminiferi bentonici del Paleocene e dell'Eocene della Sezione Posagno, *Schweiz. Paleontol. Abh.*, *97*, 85–111, 1975.

Brass, G. W., J. R. Southam, and W. H. Peterson, Warm saline bottom water in the ancient ocean, *Nature*, *296*, 620–623, 1982.

Bremer, M., and G. P. Lohmann, Evidence for primary control of the distribution of certain Atlantic Ocean benthonic foraminifera by degree of carbonate saturation, *Deep Sea Res.*, *29*, 987–998, 1982.

Broecker, W. S., and T.-H. Peng, *Tracers in the Sea*, 690 pp., Eldigio Press, Palisades, N. Y., 1982.

Butler, R. F., P. Gingerich, and E. H. Lindsay, Magnetic polarity stratigraphy and biostratigraphy of Paleocene and lower Eocene continental deposits, Clark's Fork Basin, Wyoming, *J. Geol.*, *89*, 299–316, 1981.

Butler, R. F., D. Krause, and P. Gingerich, Magnetic polarity stratigraphy and biostratigraphy of middle-late Paleocene continental deposits of south-central Montana, *J. Geol.*, *95*, 647–657, 1987.

Caralp, M. H., Impact de la matière organique dans les zones de forte productivité sur certains foraminiferes benthiques, *Oceanol. Acta*, *7*, 509–515, 1984.

Chamberlin, T. C., On a possible reversal of deep-sea circulation and its influence on geologic climates, *J. Geol.*, *14*, 363–373, 1906.

Comiso, J. C., and A. L. Gordon, Recurring polynyas over the Cosmonaut Sea and the Maud Rise, *J. Geophys. Res.*, *92*, 2819–2833, 1987.

Corfield, R. M., and J. Cartlidge, Isotopic evidence for the depth stratification of fossil and recent Globigerinina: A review, *Hist. Biol.*, *5*, 37–63, 1992.

Corfield, R. M., and N. J. Shackleton, Productivity change as a control on planktonic foraminiferal evolution after the Cretaceous/Tertiary boundary, *Hist. Biol.*, *1*, 323–343, 1988.

Corfield, R. M., J. E. Cartlidge, I. Premoli-Silva, and R. A. Housley, Oxygen and carbon isotope stratigraphy of the Paleogene and Cretaceous limestones in the Bottacione Gorge and the Contessa Highway sections, Umbria, Italy, *Terra Nova*, *3*, 414–422, 1991.

Corliss, B. H., Deep-sea benthonic foraminiferal faunal turnover near the Eocene/Oligocene boundary, *Mar. Micropaleontol.*, *6*, 367–384, 1981.

Corliss, B. H., Microhabitats of benthic foraminifera within deep-sea sediments, *Nature*, *314*, 435–438, 1985.

Corliss, B. H., and C. Chen, Morphotype patterns of Norwegian deep-sea benthic foraminifera and ecological implications, *Geology*, *16*, 716–719, 1988.

Corliss, B. H., and S. Emerson, Distribution of Rose Bengal stained deep-sea benthic foraminifera from the Nova Scotian continental margin and Gulf of Maine, *Deep Sea Res.*, *37*, 381–400, 1990.

Corliss, B. H., and L. D. Keigwin, Jr., Eocene-Oligocene paleoceanography, in *Mesozoic and Cenozoic Oceans*, *Geodyn. Ser.*, vol. 15, edited by K. Hsu, pp. 101–118, AGU, Washington, D. C., 1986.

Corliss, B. H., M.-P. Aubry, W. A. Berggren, J. M. Fenner, L. D. Keigwin, Jr., and G. Keller, The Eocene/Oligocene boundary in the deep sea, *Science*, *226*, 806–810, 1984.

Culver, S. J., Foraminifera, in *Fossil Prokaryotes and Protists*, *Stud. in Geol.*, vol. 18, edited by J. R. Lipps, pp. 169–212, University of Tennessee, Knoxville, 1987.

Cushman, J. A., Upper Cretaceous foraminifera of the Gulf Coastal region of the United States and adjacent areas, *U.S. Geol. Surv. Prof. Pap.*, *206*, 241 pp., 1946.

Dailey, D. H., Late Cretaceous and Paleocene benthic foraminifera from DSDP Site 516, Rio Grande Rise, western South Atlantic, *Initial Rep. Deep Sea Drill. Proj.*, *74*, 757–782, 1983.

Diester-Haass, L., Eocene/Oligocene paleoceanography in the Antarctic Ocean, Atlantic sector (Maud Rise, ODP Leg 113, sites 689B and 690B), *Mar. Geol.*, *100*, 249–276, 1991.

Douglas, R. G., and F. Woodruff, Deep sea benthic foraminifera, in *The Sea*, vol. 7, *The Oceanic Lithosphere*, edited by C. Emiliani, pp. 1233–1327, John Wiley, New York, 1981.

Eldholm, O., Paleogene North Atlantic magmatic-tectonic events: Environmental implications, *Mem. Soc. Geol. Ital.*, *44*, 13–28, 1990.

Ganapathy, R., Evidence for a major meteorite impact on the Earth 34 million years ago: Implication for Eocene extinctions, *Science*, *216*, 885–886, 1982*a*.

Ganapathy, R., Evidence for a major meteorite impact on the Earth 34 million years ago: Implication for the origin of North American tektites and Eocene extinctions, *Geol. Soc. Am. Spec. Publ.*, *190*, 513–516, 1982*b*.

Glass, B. P., Possible correlations between tektite events and climatic changes?, *Geol. Soc. Am. Spec. Publ.*, *190*, 251–256, 1982.

Glass, B. P., and M. J. Zwart, North American microtektites in Deep Sea Drilling Project cores from the Caribbean Sea and Gulf of Mexico, *Geol. Soc. Am. Bull.*, *90*, 595–602, 1979.

Glass, B. P., R. N. Baker, D. Storzer, and G. A. Wagner, North American microtektites from the Caribbean Sea and their fission track ages, *Earth Planet. Sci. Lett.*, *19*, 184–192, 1983.

Goldfarb, R. J., L. W. Snee, L. D. Miller, and R. J. Newberry, Rapid dewatering of the crust deduced from ages of mesothermal gold deposits, *Nature*, *354*, 296–298, 1991.

Gooday, A. J., A response by benthic foraminifera to the deposition of phytodetritus in the deep sea, *Nature*, *332*, 70–73, 1988.

Gooday, A. J., and P. J. D. Lambshead, Influence of seasonally deposited phytodetritus on benthic foraminiferal populations in the bathyal northeast Atlantic: The species response, *Mar. Ecol. Prog. Ser.*, *58*, 53–67, 1989.

Gooday, A. J., and C. M. Turley, Responses by benthic organisms to inputs of organic material to the ocean floor: A review, *Philos. Trans. R. Soc. London, Ser. A*, *331*, 119–138, 1990.

Gooday, A. J., L. A. Levin, P. Linke, and T. Heeger, The role of benthic foraminifera in deep-sea food webs and carbon cycling, in G. T. Rowe and V. Pariente, eds., *Deep-Sea Food Chains and the Global Carbon Cycle*, edited by G. T. Rowe and V. Pariente, pp. 63–91, Kluwer Academic Publishers, Dordrecht, Netherlands, 1992.

Hallock, P., I. Premoli-Silva, and A. Boersma, Similarities between planktonic and larger foraminiferal evolutionary trends through Paleogene paleoceanographic changes, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 49–64, 1991.

Haq, B. U., I. Premoli-Silva, and G. P. Lohmann, Calcareous plankton biogeographic evidence for major climatic fluctuations in the early Cenozoic Atlantic Ocean, *J. Geophys. Res.*, *82*, 3861–3876, 1977.

Haq, B. U., J. Hardenbol, and P. R. Vail, The chronology of fluctuation of sealevel since the Triassic, *Science*, *235*, 1156–1167, 1987.

Hay, W. W., Paleoceanography: A review for GSA centennial, *Geol. Soc. Am. Bull.*, *100*, 1934–1956, 1989.

Herbert, T. D., and J. L. Sarmiento, Ocean nutrient distribution and oxygenation: Limits on the formation of warm saline bottom water over the past 91 m.y., *Geology*, *19*, 702–705, 1991.

Hovan, S. A., and D. K. Rea, Paleocene/Eocene boundary changes in atmospheric and oceanic circulation: A southern hemisphere record, *Geology*, *20*, 15–18, 1992.

Kaiho, K., Uppermost Cretaceous to Paleogene bathyal benthic foraminiferal biostratigraphy of Japan and New Zealand; latest Paleocene–middle Eocene benthic foraminiferal species turnover, *Rev. Paleobiol. Spec. Vol.*, *2*, 553–559, 1988.

Kaiho, K., Global changes of Paleogene aerobic/anaerobic benthic foraminifera and deep-sea circulation, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 65–86, 1991.

Kaiho, K., A low extinction rate of intermediate-water benthic foraminifera at the Cretaceous/Tertiary boundary, *Mar. Micropaleontol.*, *18*, 229–259, 1992.

Katz, M. R., and K. G. Miller, Early Paleogene benthic foraminiferal assemblage and stable isotope composition in the Southern Ocean, ODP Leg 114, *Proc. Ocean Drill. Program Sci. Results*, *114*, 481–513, 1991.

Keigwin, L. D., Jr., and B. H. Corliss, Stable isotopes in late middle Eocene through Oligocene foraminifera, *Geol. Soc. Am. Bull.*, *97*, 335–345, 1986.

Keigwin, L. D., Jr., and G. Keller, Middle Oligocene cooling from equatorial Pacific DSDP Site 77B, *Geology*, *12*, 16–19, 1984.

Keith, M. L., Violent volcanism, stagnant oceans and some inferences regarding petroleum, strata-bound ores and mass extinctions, *Geochim. Cosmochim. Acta*, *46*, 2621–2637, 1982.

Keller, G., Biochronology and paleoclimatic implications on middle Eocene through Oligocene planktic foraminiferal faunas, *Mar. Micropaleontol.*, *7*, 474–486, 1983*a*.

Keller, G., Paleoclimatic analysis of middle Eocene through Oligocene planktic foraminiferal faunas, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *49*, 73–94, 1983*b*.

Keller, G., Stepwise mass extinctions and impact events: Late Eocene to early Oligocene, *Mar. Micropaleontol.*, *10*, 267–293, 1986.

Keller, G., Biotic turnover among benthic foraminifera across the Cretaceous/Tertiary boundary, El Kef, Tunisia, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *66*, 153–171, 1988.

Keller, G., S. L. D'Hondt, and T. Vallier, Multiple microtektite horizons in upper Eocene marine sediments: No evidence for mass extinctions, *Science*, *221*, 150–152, 1983.

Kemp. E. M., Tertiary climatic evolution and vegetation history in the southeast Indian Ocean region, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *24*, 169–208, 1978.

Kennett, J. P., Cenozoic evolution of Antarctic glaciation, the circum-Antarctic Ocean, and their impact on global paleoceanography, *J. Geophys. Res.*, *82*, 3843–3860, 1977.

Kennett, J. P., The development of planktonic biogeography in the Southern Ocean during the Cenozoic, *Mar. Micropaleontol.*, *3*, 301–345, 1978.

Kennett, J. P., Paleoceanographic changes at the Paleocene/Eocene boundary in mid-latitude ODP Site 762C, Exmouth Plateau, NW Australia (abstract), *Geol. Soc. Am. Abstr. Programs*, *23*(5), A338, 1991.

Kennett, J. P., and P. F. Barker, Latest Cretaceous to Cenozoic climate and oceanographic developments in the Weddell Sea, Antarctica: An ocean-drilling perspective, *Proc. Ocean Drill. Program Sci. Results*, *113*, 937–962, 1990.

Kennett, J. P., and N. J. Shackleton, Oxygen isotope evidence for the development of the psychrosphere 38 myr. ago, *Nature*, *260*, 513–515, 1976.

Kennett, J. P., and L. D. Stott, Proteus and Proto-Oceanus: Ancestral Paleogene oceans as revealed from Antarctic stable isotopic results, *Proc. Ocean Drill. Program Sci. Results*, *113*, 865–880, 1990.

Kennett, J. P., and L. D. Stott, Terminal Paleocene deep-sea benthic crisis: Sharp deep sea warming and paleoceanographic changes in Antarctica, *Nature*, *353*, 225–229, 1991.

Klootwijk, C. T., J. S. Gee, J. W. Peirce, and G. M. Smith, Constraints on the India-Asia convergence: Paleomagnetic results from Ninetyeast Ridge, *Proc. Ocean Drill. Program Sci. Results*, *121*, 777–882, 1991.

Knox, R. W. O., and A. C. Morton, Stratigraphical distribution of early Palaeogene pyroclastic deposits in the North Sea Basin, *Proc. Yorks. Geol. Soc.*, *44*, 355–363, 1983.

Knox, R. W. O., and A. C. Morton, The record of early Tertiary N. Atlantic volcanism in sediments of the North Sea Basin, *Geol. Soc. Spec. Publ. London*, *39*, 407–420, 1988.

Koch, P. L., J. C. Zachos, and P. D. Gingerich, Coupled isotopic change in marine and continental carbon reservoirs at the Palaeocene/Eocene boundary, *Nature*, *358*, 319–322, 1992.

Kurihara, K., and J. P. Kennett, Bathymetric migration of deep-sea benthic foraminifera in the southwest Pacific during the Neogene, *J. Foraminiferal Res.*, *18*, 75–83, 1988.

Kyte, F. T., M. Leinen, G. Ross Heath, and L. Zhou, Elemental geochemistry of Core LL44-GPC3 and a model for the Cenozoic sedimentation history of the central North Pacific, *Geochim. Cosmochim. Acta*, in press, 1992.

Lambshead, P. J. D., and A. J. Gooday, The impact of seasonally deposited phytodetritus on epifaunal and shallow infaunal benthic foraminiferal populations in the bathyal northeast Atlantic: The assemblage response, *Deep Sea Res.*, *37*, 1263–1283, 1990.

Leinen, M., The origin of paleochemical signatures in North Pacific pelagic clays: Partitioning experiments, *Geochim. Cosmochim. Acta*, *51*, 305–319, 1987.

Leinen, M., and D. Stakes, Metal accumulation rates in the central equatorial Pacific during Cenozoic times, *Geol. Soc. Am. Bull.*, *90*, 357–375, 1979.

Lenotre, N., H. Chamley, and M. Hoffert, Clay stratigraphy at DSDP sites 576 and 578, DSDP Leg 86 (western North Pacific), *Initial Rep. Deep Sea Drill. Proj.*, *86*, 605–646, 1985.

Lohmann, G. P., Abyssal benthonic foraminifera as hydrographic indicators in the western South Atlantic Ocean, *J. Foraminiferal Res.*, *8*, 6–34, 1978.

Lu, G., and G. Keller, Climatic and oceanographic events across the Paleocene-Eocene transition in the Antarctic Indian Ocean: Inference from planktic foraminifera, *Mar. Micropaleontol.*, in press, 1992.

Lutze, G. F., and W. T. Coulbourn, Recent benthic foraminifera from the continental margin of northwest Africa: Community structure and distribution, *Mar. Micropaleontol.*, *8*, 361–401, 1984.

Mackensen, A., and W. A. Berggren, Paleogene benthic foraminifers from the southern Indian Ocean (Kerguelen Plateau): Biostratigraphy and paleoecology, *Proc. Ocean Drill. Program Sci. Results*, *120*, 603–630, 1992.

MacLeod, N., Effects of late Eocene impacts on planktonic foraminifera, *Geol. Soc. Am. Spec. Publ.*, *247*, 595–606, 1990.

Manabe, S., and K. Bryan, CO_2-induced change in a coupled ocean-atmosphere system and its paleoclimatic implications, *J. Geophys. Res.*, *90*, 11,689–11,707, 1985.

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

McGowran, B., Late Eocene perturbations: Foraminiferal biofacies and evolutionary overturn, southern Australia, *Paleoceanography*, *2*, 715–727, 1987.

McGowran, B., Silica burp in the Eocene ocean, *Geology*, *17*, 857–860, 1989.

McGowran, B., Fifty million years ago, *Am. Sci.*, *78*, 30–39, 1990.

McGowran, B., Evolution and environment in the early Paleogene, *Mem. Geol. Soc. India*, *20*, 21–53, 1991.

Mead, G. A., and D. A. Hodell, Late Eocene to early Oligocene vertical oxygen isotope gradients in the South Atlantic: Implications for warm saline deep water, *Paleoceanography*, in press, 1992.

Mercer, J. H., Cenozoic glaciation in the southern hemisphere, *Annu. Rev. Earth Planet. Sci.*, *11*, 99–132, 1982.

Miller, K. G., Eocene-Oligocene paleoceanography in the deep Bay of Biscay, *Mar. Micropaleontol.*, *7*, 403–440, 1983.

Miller, K. G., and R. G. Fairbanks, Oligocene-Miocene global carbon and abyssal circulation changes, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to*

Present, Geophys. Monogr. Ser., vol. 32, edited by E. Sundquist and W. S. Broecker, pp. 469–486, AGU, Washington, D. C., 1985.

Miller, K. G., and M. Katz, Oligocene to Miocene benthic foraminiferal and abyssal circulation changes in the North Atlantic, *Micropaleontology*, *33*, 97–149, 1987.

Miller, K. G., and E. Thomas, Late Eocene to Oligocene benthic foraminiferal isotopic record, Site 574 Equatorial Pacific, *Initial Rep. Deep Sea Drill. Proj.*, *85*, 771–777, 1985.

Miller, K. G., W. B. Curry, and D. R. Ostermann, Late Paleogene (Eocene to Oligocene) benthic foraminiferal oceanography of the Goban Spur region DSDP Leg 80, *Initial Rep. Deep Sea Drill. Proj.*, *80*, 505–538, 1984.

Miller, K. G., R. G. Fairbanks, and G. S. Mountain, Tertiary isotope synthesis, sea level history, and continental margin erosion, *Paleoceanography*, *2*, 1–20, 1987*a*.

Miller, K. G., T. R. Janecek, M. R. Katz, and D. J. Keil, Abyssal circulation and benthic foraminiferal changes near the Paleocene/Eocene boundary, *Paleoceanography*, *2*, 741–761, 1987*b*.

Montanari, A., Geochronology of the terminal Eocene impacts: An update, *Geol. Soc. Am. Spec. Publ.*, *247*, 607–616, 1990.

Mueller-Merz, E., and H. Oberhaensli, Eocene bathyal and abyssal benthic foraminifera from a South Atlantic transect at 20°–30°S, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 117–172, 1991.

Murray, J. W., J. F. Weston, C. A. Haddon, and A. D. J. Powell, Miocene to Recent bottom water masses of the northeast Atlantic: An analysis of benthic foraminifera, in North Atlantic Palaeoceanography, *Geol. Soc. Spec. Publ. London*, *21*, 219–230, 1986.

Niensted, J. C., and A. J. Arnold, The distribution of benthic foraminifera on seamounts near the East Pacific Rise, *J. Foraminiferal Res.*, *18*, 237–249, 1988.

Nomura, R., Paleoceanography of upper Maestrichtian to Eocene benthic foraminiferal assemblages at ODP sites 752, 753 and 754, eastern Indian Ocean, *Proc. Ocean Drill. Program Sci. Results*, *121*, 3–30, 1991*a*.

Nomura, R., Oligocene to Pleistocene benthic foraminifer assemblages at sites 754 and 756, eastern Indian Ocean, *Proc. Ocean Drill. Program Sci. Results*, *121*, 31–76, 1991*b*.

Nomura, R., K. Seto, and N. Niitsuma, Late Cenozoic deep-sea benthic foraminiferal changes and isotopic records in the eastern Indian Ocean, in *Fourth International Symposium on Benthic Foraminifera*, edited by T. Saito, Yamagata University, Yamagata, Japan, in press, 1992.

Oberhänsli, H., E. Mueller-Merz, and R. Oberhänsli, Eocene paleoceanographic evolution at 20°–30°S in the Atlantic Ocean, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 173–216, 1991.

O'Connell, S., Variations in Upper Cretaceous and Cenozoic calcium carbonate percentages, Maud Rise, Weddell Sea, Antarctica, *Proc. Ocean Drill. Program Sci. Results*, *113*, 971–984, 1990.

Olivarez, A. M., and R. M. Owen, Plate tectonic reorganizations: Implications regarding the formation of hydrothermal ore deposits, *Mar. Min.*, *14*, 123–138, 1989.

Ottens, J. J., and A. J. Nederbragt, Planktic foraminiferal diversity as indicator of ocean environments, *Mar. Micropaleontol.*, *19*, 13–28, 1992.

Owen, R. M., and D. K. Rea, Sea floor hydrothermal activity links climate to tectonics: The Eocene CO_2 greenhouse, *Science*, *227*, 166–169, 1985.

Pak, D. K., and K. G. Miller, Late Paleocene to early Eocene benthic foraminiferal stable isotopes and assemblages: Implications for deepwater circulation, *Paleoceanography*, in press, 1992.

Pak, D. K., K. G. Miller, and J. D. Wright, Early Paleogene deep and intermediate water sources: Evidence from benthic foraminiferal and isotopic changes (abstract), *Geol. Soc. Am. Abstr. Programs*, *23*(5), A141–A142, 1991.

Poore, R. Z., and R. K. Matthews, Late Eocene–Oligocene oxygen and carbon isotopic record from the South Atlantic DSDP Site 522, *Initial Rep. Deep Sea Drill. Proj.*, *73*, 725–735, 1984.

Pospichal, J. J., and S. W. Wise, Jr., Paleocene to middle Eocene calcareous nannofossils of ODP sites 689 and 690, Maud Rise, Weddell Sea, *Proc. Ocean Drill. Program Sci. Results*, *113*, 613–638, 1990.

Premoli-Silva, I., and A. Boersma, Atlantic Eocene planktonic foraminiferal historical biogeographic and paleohydrologic indices, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *67*, 315–356, 1984.

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

Prothero, D. A., and W. A. Berggren (Eds.), *Eocene-Oligocene Climatic and Biotic Evolution*, 568 pp., Princeton University Press, Princeton, N. J., 1992.

Railsback, L. B., Influence of changing deep ocean circulation on the Phanerozoic oxygen isotopic record, *Geochim. Cosmochim. Acta*, *54*, 1501–1509, 1990.

Railsback, L. B., T. F. Anderson, S. C. Ackerly, and J. L. Cisne, Paleoceanographic modeling of temperature-salinity profiles from stable isotope data, *Paleoceanography*, *4*, 585–591, 1989.

Raup, D. M., and J. J. Sepkoski, Periodic extinction of families and genera, *Science*, *231*, 833–836, 1986.

Rea, D. K., J. C. Zachos, R. M. Owen, and D. Gingerich, Global change at the Paleocene-Eocene boundary: Climatic and evolutionary consequences of tectonic events, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *79*, 117–128, 1990.

Robert, C., and H. Chamley, Development of early Eocene warm climates, as inferred from clay mineral variations in oceanic sediments, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *89*, 315–332, 1991.

Robert, C., and J. P. Kennett, Paleocene and Eocene kaolinite distribution in the South Atlantic and Southern Ocean: Antarctic climatic and paleoceanographic implications, *Mar. Geol.*, *103*, 99–110, 1992.

Robert, C., and H. Maillot, Paleoenvironments in the Weddell Sea area and Antarctic climates as deduced from clay mineral associations and geochemical data, ODP Leg 113, *Proc. Ocean Drill. Program Sci. Results*, *113*, 51–70, 1990.

Roberts, D. G., A. C. Morton, and J. Backman, Late Paleocene-Eocene volcanic events in the northern North Atlantic Ocean, *Initial Rep. Deep Sea Drill. Proj.*, *81*, 913–923, 1984.

Romein, A. J. T., Lineages in early Paleogene calcareous nannoplankton, *Utrecht Micropaleontol. Bull.*, *22*, 231 pp., 1979.

Rosoff, D. B., and B. H. Corliss, An analysis of Recent deep-sea benthic foraminiferal morphotypes from the Norwegian and Greenland seas, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *91*, 13–20, 1992.

Sanders, H. L., Marine benthic diversity: A comparative study, *Am. Nat.*, *102*, 243–282, 1968.

Savin, S. M., The history of the Earth's surface temperature during the past 100 million years, *Annu. Rev. Earth Planet. Sci.*, *5*, 319–344, 1977.

Savin, S. M., R. G. Douglas, G. Keller, J. S. Killingley, L. Shaughnessy, M. A. Sommer, E. Vincent, and F. Woodruff, Miocene benthic foraminiferal isotope records: A synthesis, *Mar. Micropaleontol.*, *6*, 423–450, 1981.

Schmidt, K.-H., Tertiary palaeoclimatic history of the southeastern Colorado Plateau, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *86*, 283–296, 1991.

Schnitker, D., West Atlantic abyssal circulation during the past 120,000 years, *Nature*, *248*, 385–387, 1974.

Schnitker, D., The deep waters of the western North Atlantic during the past 24,000 years, and reinitiation of the Western Boundary Undercurrent, *Mar. Micropaleontol.*, *4*, 265–280, 1979*a*.

Schnitker, D., Cenozoic deep water benthic foraminifers, Bay of Biscay, *Initial Rep. Deep Sea Drill. Proj.*, *48*, 377–414, 1979*b*.

Schnitker, D., North-east Atlantic Neogene benthic foraminiferal faunas: Tracers of deepwater palaeoceanography, in North Atlantic Palaeoceanography, *Geol. Soc. Spec. Publ. London*, *21*, 191–204, 1986.

Schroeder-Adams, C. J., Eocene to Recent benthic foraminifer assemblages from the Kerguelen Plateau (southern Indian Ocean), *Proc. Ocean Drill. Program Sci. Results*, *119*, 611–630, 1991.

Sen Gupta, B. K., R. F. Lee, and S. M. Mallory, Upwelling and an unusual assemblage of benthic foraminifera on the northern Florida continental slope, *J. Paleontol.*, *55*, 853–857, 1981.

Shackleton, N. J., Oxygen isotopic evidence for Cenozoic climate change, in *Fossils and Climate*, edited by P. Brenchley, pp. 27–34, John Wiley, New York, 1984.

Shackleton, N. J., Paleogene stable isotope events, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *57*, 91–102, 1986.

Shackleton, N. J., The carbon isotope history of the Cenozoic, in *Petroleum Source Rocks*, edited by J. Brooks and A. J. Fleet, pp. 427–438, Blackwell Scientific, Boston, 1987.

Shackleton, N. J., and A. Boersma, The climate of the Eocene ocean, *J. Geol. Soc. London*, *138*, 153–157, 1981.

Shackleton, N. J., and M. A. Hall, Carbon isotope data from Leg 74 sediments, *Initial Rep. Deep Sea Drill. Proj.*, *74*, 613–619, 1984.

Shackleton, N. J., and M. A. Hall, Carbon isotope stratigraphy of bulk sediments, ODP sites 689 and 690, Maud Rise, Antarctica, *Proc. Ocean Drill. Program Sci. Results*, *113*, 985–989, 1990.

Shackleton, N. J., and J. P. Kennett, Palaeotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analyses in DSDP sites 277, 279 and 281, *Initial Rep. Deep Sea Drill. Proj.*, *29*, 743–755, 1975.

Shackleton, N. J., M. A. Hall, and A. Boersma, Oxygen and carbon isotope data from Leg 74 foraminifers, *Initial Rep. Deep Sea Drill. Proj.*, *74*, 599–612, 1984*a*.

Shackleton, N. J., et al., Accumulation rates in Leg 74 sediments, *Initial Rep. Deep Sea Drill. Proj.*, *74*, 621–637, 1984*b*.

Shackleton, N. J., M. A. Hall, and U. Bleil, Carbon isotope stratigraphy, Site 577, *Initial Rep. Deep Sea Drill. Proj.*, *86*, 503–511, 1985.

Signor, P., and J. Lipps, Sampling bias, gradual extinction patterns, and catastrophes in the fossil record, *Geol. Soc. Am. Spec. Publ.*, *19*, 291–296, 1982.

Smart, C. W., Ecological controls on patterns of speciation and extinction in deep-sea benthic foraminifera, M.Phil. report, Univ. of Southampton, Southampton, United Kingdom, 1991.

Spezzaferri, S., and I. Premoli-Silva, Oligocene planktonic foraminiferal biostratigraphy and paleoclimatic interpretation from Hole 538A, DSDP Leg 77, Gulf of Mexico, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *83*, 217–263, 1991.

Stott, L. D., and J. P. Kennett, Antarctic Paleogene planktonic foraminiferal biostratigraphy: ODP Leg 113, sites 689 and 690, *Proc. Ocean Drill. Program Sci. Results*, *113*, 549–570, 1990.

Stott, L. D., J. P. Kennett, N. J. Shackleton, and R. M. Corfield, The evolution of Antarctic surface waters during the Paleogene: Inferences from the stable isotopic composition of planktonic foraminifera, ODP Leg 113, *Proc. Ocean Drill. Program Sci. Results*, *113*, 849–864, 1990.

Streeter, S. S., Bottom water and benthonic foraminifera in the North Atlantic: Glacial-interglacial contrasts, *Quat. Res.*, *3*, 131–141, 1973.

Streeter, S. S., and N. J. Shackleton, Paleocirculation of the deep North Atlantic: 150,000 year record of benthic foraminifera and oxygen-18, *Science*, *203*, 168–171, 1979.

Thiel, H., O. Pfannkuche, G. Schriever, K. Lochte, A. J. Gooday, C. Hemleben, R. F. G. Mantoura, C. M. Turley, J. W. Patching, and F. Riemann, Phytodetritus on the deep-sea floor in a central oceanic region of the northeast Atlantic, *Biol. Oceanogr.*, *6*, 203–239, 1988.

Thierstein, H. R., Inventory of paleoproductivity records: The Mid-Cretaceous enigma, in *Productivity of the Oceans: Past and Present*, edited by W. H. Berger, V. S. Smetacek, and G. Wefer, pp. 355–375, John Wiley, New York, 1989.

Thomas, E., Late Eocene to Recent deep-sea benthic foraminifers from the central equatorial Pacific Ocean, *Initial Rep. Deep Sea Drill. Proj.*, *85*, 655–679, 1985.

Thomas, E., Changes in composition of Neogene benthic foraminiferal faunas in equatorial Pacific and North Atlantic, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *53*, 47–61, 1986*a*.

Thomas, E., Early to middle Miocene benthic foraminiferal faunas from DSDP sites 608 and 610, North Atlantic, in North Atlantic, Palaeoceanography, *Geol. Soc. Spec. Publ. London*, *21*, 205–218, 1986*b*.

Thomas, E., Development of Cenozoic deep-sea benthic foraminiferal faunas in Antarctic waters, in Origins and Evolution of Antarctic Biota, *Geol. Soc. Spec. Publ. London*, *47*, 283–296, 1989.

Thomas, E., Late Cretaceous through Neogene deep-sea benthic foraminifers, Maud Rise, Weddell Sea, Antarctica, *Proc. Ocean Drill. Program Sci. Results*, *113*, 571–594, 1990*a*.

Thomas, E., Late Cretaceous–early Eocene mass extinctions in the deep sea, in *Global Catastrophes*, *Geol. Soc. Am. Spec. Publ.*, *247*, 481–496, 1990*b*.

Thomas, E., The latest Paleocene mass extinction of deep-sea benthic foraminifera: Result of global change (abstract), *Geol. Soc. Am. Abstr. Programs*, *23*(5), A141, 1991.

Thomas, E., Middle Eocene–late Oligocene bathyal benthic foraminifera (Weddell Sea): Faunal changes and implications for ocean circulation, in *Eocene-Oligocene Climatic and Biotic Evolution*, edited by D. A. Prothero and W. A. Berggren, pp. 245–271, Princeton University Press, Princeton, N. J., 1992.

Thomas, E., and J. C. Varekamp, Did volcanic CO_2 emissions cause the rapid global change at the end of the Paleocene?, paper presented at Chapman Conference on Climate, Volcanism and Global Change, AGU, Hilo, Hawaii, March 23–27, 1992.

Thomas, E., and E. Vincent, Major changes in benthic foraminifera in the equatorial Pacific before the middle Miocene polar cooling, *Geology*, *15*, 1035–1039, 1987.

Thomas, E., and E. Vincent, Early to middle Miocene deep-sea benthic foraminifera in the equatorial Pacific, *Rev. Paleobiol. Spec. Vol.*, *2*, 583–588, 1988.

Thomas, E., N. J. Shackleton, and M. A. Hall, Carbon isotope stratigraphy of Paleogene bulk sediments, Hole 762C (Exmouth Plateau, eastern Indian Ocean), *Proc. Ocean Drill. Program Sci. Results*, *122*, 897–901, 1992.

Tjalsma, R. C., and G. P. Lohmann, Paleocene-Eocene bathyal and abyssal benthic foraminifera from the Atlantic Ocean, *Micropaleontology Spec. Publ.*, *4*, 94 pp., 1983.

van Andel, T. H., Mesozoic-Cenozoic calcite compensation depth and the global distribution of carbonate sediments, *Earth Planet. Sci. Lett.*, *26*, 187–194, 1975.

van der Zwaan, G. J., F. J. Jorissen, and W. J. Zachariasse (Eds.), Approaches to paleoproductivity reconstructions, *Mar. Micropaleontol.*, *19*, 1–180, 1992.

van Morkhoven, F. P. C., W. A. Berggren, and A. S. Edwards, *Cenozoic Cosmopolitan Deep-Water Benthic Foraminifera*, 421 pp., Elf-Aquitaine, Pau, France, 1986.

Vincent, E., and J. S. Killingley, Oxygen and carbon isotope record for the early and middle Miocene in the central equatorial Pacific (DSDP Leg 85), *Initial Rep. Deep Sea Drill. Proj.*, *85*, 749–770, 1985.

Webb, P.-N., Upper Cretaceous–Paleocene foraminifera from Site 208, Lord Howe Rise, Tasman Sea, DSDP Leg 21, *Initial Rep. Deep Sea Drill. Proj.*, *21*, 541–573, 1973.

Webb, P.-N., The Cenozoic history of Antarctica and its global impact, *Antarct. Sci.*, *2*, 3–21, 1990.

Wei, W., Evidence for an earliest Oligocene abrupt cooling in the surface waters of the Southern Ocean, *Geology*, *19*, 780–783, 1991.

White, R. S., Igneous outbursts and mass extinctions, *Eos Trans. AGU*, *70*, 1480, 1490–1491, 1989.

White, R. S., and D. Mackenzie, Magmatism at rift zones: The generation of volcanic continental margins and flood basalts, *J. Geophys. Res.*, *94*, 7685–7729, 1989.

Widmark, J. G. V., Upper Cretaceous–lower Tertiary deep-sea benthic foraminifera from the Walvis Ridge, South Atlantic Ocean: Taxonomy, paleobiogeography, and paleoecology, *Abstr. Uppsala Diss. Sci.*, *283*, 1–289, 1990.

Widmark, J. G. V., and B. Malmgren, Benthic foraminiferal changes across the Cretaceous-Tertiary boundary in the deep sea: DSDP sites 525, 527 and 465, *J. Foraminiferal Res.*, *22*, 81–113, 1992.

Williams, C. A., An oceanwide view of Palaeogene plate tectonics, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *57*, 3–25, 1986.

Wing, S. L., A new basis for recognizing the Paleocene/Eocene boundary in western interior North America, *Science*, *226*, 439–441, 1984.

Wing, S. L., T. M. Bown, and J. D. Obradovich, Early Eocene biotic and climatic change in interior western America, *Geology*, *19*, 1189–1192, 1991.

Wise, S. W., J. R. Breza, D. M. Harwood, and W. Wei, Paleogene glacial history of Antarctica, in *Controversies in Modern Geology*, pp. 133–171, Academic, San Diego, Calif., 1991.

Wise, S. W., Jr., J. R. Breza, D. M. Harwood, W. Wei, and J. C. Zachos, Paleogene glacial history of Antarctica in the light of ODP Leg 120 drilling results, *Proc. Ocean Drill. Program Sci. Results*, *120*, 1001–1030, 1992.

Wolfe, J. A., A paleobotanical interpretation of Tertiary climates in the northern hemisphere, *Am. Sci.*, *66*, 694–703, 1978.

Wolfe, J. A., and R. Z. Poore, Tertiary marine and non-marine climatic trends, in *Climate in Earth History*, pp. 154–158, National Academy Press, Washington, D. C., 1982.

Woodruff, F., Changes in Miocene deep sea benthic foraminiferal distribution in the Pacific Ocean: Relationship to paleoceanography, *Mem. Geol. Soc. Am.*, *163*, 131–176, 1985.

Woodruff, F., and R. G. Douglas, Response of deep sea benthic foraminifera to Miocene paleoclimatic events, *Mar. Micropaleontol.*, *6*, 617–632, 1981.

Woodruff, F., and S. M. Savin, d^{13}C values of Miocene Pacific benthic foraminifera: Correlations with sea level and productivity, *Geology*, *13*, 119–122, 1985.

Woodruff, F., and S. M. Savin, Miocene deep water oceanography, *Paleoceanography*, *4*, 87–140, 1989.

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

Wright, J. D., and K. G. Miller, Southern Ocean dominance of late Eocene to Oligocene deep-water circulation (abstract), *Eos Trans. AGU*, *73*, 171, 1992.

Zachos, J., W. A. Berggren, M.-P. Aubry, and A. Mackensen, Isotope and trace element geochemistry of Eocene and Oligocene foraminifers from Site 748, Kerguelen Plateau, *Proc. Ocean Drill. Program Sci. Results*, *120*, 839–854, 1992*a*.

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

Zachos, J., D. Rea, K. Seto, R. Nomura, and N. Niitsuma, Paleogene and early Neogene deepwater paleoceanography of the Indian Ocean as determined from benthic foraminifera stable isotope records, in *The Indian Ocean: A Synthesis of Results From the Ocean Drilling Program*, *Geophys. Monogr. Ser.*, edited by R. A. Duncan et al., AGU, Washington, D. C., in press, 1992*c*.

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