Cenozoic Deep-sea Benthic Foraminifera: 
A Tale of Three Turnovers

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ABSTRACT: Recently published syntheses of the taxonomy and biostratigraphy of Cenozoic bathyal (200-2000 m) and abyssal (> 2000 m) calcareous benthic foraminifera provide a framework for evaluating benthic faunal abundance changes and taxonomic turnovers. During the Cenozoic, there were several periods of taxonomic stability in the deep sea. There were few global first and last occurrences in the deep sea across the Cretaceous/Paleogene boundary, during the Paleocene, and in the late Neogene. There were at least three intervals of increased taxonomic turnover and faunal abundance change in deep-water benthic foraminifera: 1) a dramatic extinction in the latest Paleocene (late Biochron P16a), followed by an early Eocene period of rapid recolonization; 2) a faunal turnover in the late middle Eocene to earliest Oligocene (Biochron P14-P18); and 3) a faunal turnover in the late early to early middle Miocene (Biochron N6 to N12). We briefly review the nature and possible causes of these three faunal changes and provide new data on the middle Eocene-earliest Oligocene faunal change. The three faunal changes were triggered by different mechanisms. We attribute the latest Paleocene "crisis" to an abrupt, transient warming related to a change in deep-water source regions. The late middle Eocene-early Oligocene taxonomic turnover was relatively minor, although a dramatic decrease in the abundance of Nautilides truempyi occurred throughout the deep sea in the late middle Eocene (Biochron P14). We directly link this abundance change to a cooling of deep water inferred from the benthic foraminiferal δ13C record. The early-middle Miocene faunal turnover and abundance changes were not directly related to deep-water temperature history, although they may have been related to changes in deep-water sources and/or surface ocean productivity.

INTRODUCTION

Pioneering studies of Cenozoic deep-water (> 200 m) benthic foraminifera were restricted to modern taxa from the deep sea (e.g., Brady, 1984) or Cenozoic taxa from uplifted sections primarily in the Caribbean region (e.g., Cushman and Jarvis, 1948; Beckmann, 1954, 1960; see Douglas and Woodruff, 1981 for a historical review). These studies were largely taxonomic and regional in scope. Recovery of deep-water sections by the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Program (ODP) provided a global array suitable for benthic foraminiferal biostratigraphic, biostratigraphic, and paleobathymetric studies (e.g., Douglas, 1973; Douglas and Woodruff, 1981; Tjalsma and Lohmann, 1983; Miller, 1983; Tjalsma, 1985; Boersma, 1985, 1986; Thomas, 1985, 1986a; Woodruff, 1985; Miller and Katz, 1987a, b). Despite these advances, comparisons among previous studies were hampered by differing taxonomies. Examination of museum collections revealed the need to return to basics for consistent taxonomy (van Morkhoven et al., 1986). In addition, DSDP/ODP has primarily sampled lower bathyal to abyssal locations (> 1000 m, see below), and therefore, our understanding of the upper and middle bathyal zones (200-1000 m) is limited.

Van Morkhoven et al. (1986) provided a revised taxonomic framework and synthesized distributional data for deep-water, predominantly calcareous benthic foraminifera from the bathyal and abyssal realms. Berggren and Miller (1989) integrated this benthic foraminiferal database with magnetostratigraphy and planktonic biostratigraphy and erected 14 bathyal and 12 abyssal zones for Cenozoic deep-water benthic foraminifera. In addition, van Morkhoven et al. (1986) and Berggren and Miller (1989) evaluated depth changes of taxa through time. They used calibrations of benthic foraminiferal ranges and abundances against independent paleobathymetric estimates (primarily derived from simple age-subsidence models or "backtracking"; e.g., Tjalsma and Lohmann, 1983; Miller and Katz, 1987a), recognizing the following bathymetric divisions: neritic < 200 m

bathyal 200-2000 m
upper bathyal 200-600 m
middle bathyal 600-1000 m
lower bathyal 1000-2000 m
abyssal > 2000 m
upper abyssal 2000-3000 m
lower abyssal > 3000 m

The studies of van Morkhoven et al. (1986) and Berggren and Miller (1989) showed that: 1) many bathyal-abyssal taxa invaded the deep sea from neritic habitats (e.g., the migrations of Uvigerina, Planulina, Melonis, Siphonina, and Sphaerooidina; Figure 1); 2) many taxa migrated within the bathyal-abyssal zones through time; and 3) depth-dependent ecophenotypes must be considered in species concepts. For example, the Cibicidoides simplex-costatus group probably represents a neritic-upper bathyal morphotype of the lower bathyal-abyssal C. mexicanus group (van Morkhoven et al., 1986).

Previous studies indicate there were at least three major intervals of change in Cenozoic deep-sea benthic foraminifera...
THE PALEOCENE/EOCENE TURNOVER: AN ABRUPT EVENT TRIGGERS A MAJOR EXTINCTION

Unlike planktonic organisms, deep-sea benthic foraminifera experienced little change across the Cretaceous/Paleocene boundary (Figure 1; Beckmann, 1968; Dailey, 1983; Tjalsma and Lohmann, 1983; Thomas, in press a). A major benthic foraminiferal "crisis" occurred near the end of the Paleocene in the Atlantic (Figure 1; Tjalsma, 1976; Schnitker, 1979; Tjalsma and Lohmann, 1983), Caribbean (Tjalsma and Lohmann, 1983), Pacific (Miller et al., 1987a), Southern (Thomas, 1988, 1990; Katz and Miller, 1991; Mackensen and Berggren, in press), and Indian Oceans (Sigal, 1974; Vincens et al., 1974; Nomura, in press). The latest Paleocene "crisis" consisted of the extinction of many benthal and abyssal benthic foraminiferal taxa (Figure 1).

The timing and synchrony of this extinction event among locations is still being determined. Recent studies indicate that the benthic foraminiferal extinction event probably predates the Paleocene/Eocene boundary (see discussion below); this boundary is recognized either by the last occurrence of the planktonic foraminifera Morozovella velascoensis (= base of Zone P6b; 57.8 Ma on the time scale of Berggren et al., 1985) or the middle part of Zone N10 (Aubry et al., 1988; 57.0 Ma using the time scale of Berggren et al., 1985). We use the Berggren et al. (1985) time scale in this contribution and the Paleocene/Eocene boundary of Aubry et al. (1988), but note that the time scale for this interval will be shifted by up to 2 m.y. younger (Berggren and Kent, 1989). Nevertheless, the magnetostratigraphic relationships reported here will remain unchanged, and only the numerical ages will be different.

Only about one half of deep-water benthic foraminiferal species survived the latest Paleocene faunal turnover that occurred during Biochrons P5-P6a and NP9. Fifty percent of the Paleocene benthic foraminiferal taxa disappeared between samples in Zones P5 and P6a from Atlantic and Caribbean locations (Tjalsma and Lohmann, 1983). At Pacific Deep Sea Drilling Project (DSDP) Site S77, about 50 percent of the benthic foraminiferal species disappeared in the middle of Chron C24r and Biochron P6a (Miller et al., 1987a); closely spaced sampling reveals that the extinctions occurred 1 m below the LO of M. velascoensis at Site S77, with an age estimate of ~58.0 Ma (assuming a 4.1 m/my sedimentation rate; Pak and Miller, in press). In subtropical Ocean Drilling Program (ODP) Leg 114 sites, the extinctions occurred prior to the LO of the calcareous nannofossil Fasciiculithus (Katz and Miller, 1991), suggesting that the extinctions were older than ~57.4 Ma (using the Berggren et al., 1985 time scale). Nomura (in press) established that the extinctions occurred near the end of Biochron CP8 (= NP9, >57.8 Ma) at Indian Ocean Site 752.

Thomas (1990) demonstrated that the turnover was extremely rapid at Maud Rise Site 690, occurring within a 3,000 y interval (Thomas, personal communication, 1991). At this site, the extinctions occurred in the latest Paleocene in Chron C24r, in Biochron CP8 (= NP9; prior to the FO T. bramletti, the nominate taxon for Zones NP10; Pospichal and Wise, 1990), and prior to the LO of Fasciiculithus. Thomas et al. (1990) estimated the age of the extinction as ~57.5 Ma, which would correlate with Biochron CP9 (= NP10), the early Eocene of Berggren et al. (1985), and the latest Paleocene of Aubry et al. (1988). There is an apparent biochronologic age difference between the timing of the extinctions based on biostratigraphy at the Maud Rise site versus Pacific and Indian sites (i.e., ~57.5 Ma at Site 690 versus older than ~57.8 Ma at Sites S77 and 752). This emphasizes uncertainties in high-latitude correlations. We conclude that the data available are not sufficient to document unequivocally that the extinctions were synchronous or diachronous among basins.

Synthesizing our current understanding of the timing of this event, we can firmly state that the extinctions occurred:
1) in Chron C24r;
2) prior to the LO of Fasciiculithus;
3) in the latest Paleocene sensu Aubry et al. (1988).

Considering the uncertainties in the correlation of Site 690 to low latitudes, we can also estimate that the extinctions occurred:
1) in Biochron P6a (within the concurrent range of M. velascoensis and M. subbotinii);
2) in Biochron NP9 (= CP8);
3) in the latest Paleocene sense Berggren et al. (1985); 4) at about 58.0 Ma on the time scale of Berggren et al. (1985), which will be about 2 my. younger in the revised time scale (Berggren and Kent, 1989; Cande and Kent, MS in prep.).

Abundance changes preceded and accompanied the taxonomic turnover. Relative abundance changes occurred over several million years at Pacific Site 577, beginning before the taxonomic turnover. A Bulimina beaumonti-Neoepipondos hillebrandti-Aragonia ozezanensis-Spiwallina beccariformis-dominated biotas was replaced by a Nautilites truempyi-dominated biotas in the late Paleocene at Site 577 (Miller et al., 1987a). The Leg 114 depth transect (800-3000 m paleodepth) of the subantarctic showed that a Paleocene Spiwallina beccariformis-dominated biotas was replaced by various depth-related biotas during the latest Paleocene. Thomas (1990) documented similar faunal abundance changes at Maud Rise Site 690.

A new wave of taxa gradually filled niches vacated by the latest Paleocene extinction event (Figure 1; Tjalsma and Lohmann, 1983). This benthic radiation was staggered over several million years (Tjalsma and Lohmann, 1983; Miller et al., 1987a; Katz and Miller, 1991). Among the taxa which initially appeared in the latest Paleocene to early Eocene were Abyssmina poagi, Alabamina dissonata, Aragonia aragonensis, Clinepertina spp., Bulimina impendens, Bulimina grata, and Turritina robertsi. These taxa became abundant in the Pacific, Atlantic, and Southern Oceans, where together with Nautilites truempyi, they dominated early-middle Eocene benthic biotas (Tjalsma and Lohmann, 1983; Miller et al., 1987a; Katz and Miller, 1991).

The extinction event and biotas changes affected both bathyal and abyssal faunas. Tjalsma and Lohmann (1983) recorded the progressive restriction of the S. beccariformis-dominated biotas to upper abyssal and then lower bathyal depths during the late Paleocene of the Atlantic-Caribbean. In addition to the extinctions in the lower bathyal and abyssal zones, an extinction event apparently occurred in the upper-middle bathyal realm, characterized by the disappearance of Angulogavelinella avnimelchei (Berggren and Miller, 1989; Figure 1).

The benthic foraminiferal extinction event may have been caused by deep-water warming, a change in food supply (surface ocean productivity), or changing deep-water source regions (e.g., Miller et al., 1987a; Thomas, 1990; Katz and Miller, 1991). We briefly evaluate each potential cause of the latest Paleocene crisis.

Food Supply. Miller et al. (1987a) and Thomas (1990) argued against a change in surface ocean productivity as a cause of the benthic extinction. The benthic extinctions occurred well after the start of a drop in global δ13C values, which was previously inferred to represent a change in surface ocean productivity (Shackleton et al., 1985). In addition, there apparently was little change in the vertical δ13C difference between phosphate-free surface waters and mean deep water, which weakens the case for a change in productivity (Miller et al., 1987a). Corfield and Cartledge (1990), on the other hand, argued that there was a change in the vertical δ13C gradient in this interval, suggesting a change in surface ocean productivity. Nevertheless, we caution that stable isotope data alone cannot be used to prove or disprove a global change in surface ocean productivity, and an evaluation of global sedimentary fluxes is needed. We conclude that it is possible that the extinctions were caused by a reduced food supply, although we regard it as unlikely.

Temperature change. The major extinction event occurred during a general warming of deep water between approximate-ly 60 and 56 Ma (Miller et al., 1987a). Based on their fairly coarse sampling interval (~0.5-0.5 m.y. sampling), Katz and Miller (1991) found the relationship between extinctions and the δ18O record to be unconvincing. Miller et al. (1987a) questioned why bottom-water warming should cause extinctions primarily among the shallower, presumably more thermophilic taxa (the S. beccariformis biotas) rather than the deeper taxa. However, Kennett and Stott (1991) recently presented detailed (4-8 k.y. sampling interval) stable isotope data across this interval from Site 690 that suggests that a rapid, transient warming occurred at the same time as the extinction event. Their more detailed data strongly suggest that an abrupt change in deep-water conditions (including an increase in deep-water temperatures) was responsible for the extinction event.

Changing deep-water supply. The extinction event may be linked to a change in deep-water source regions. Interbasinal carbon isotope comparisons (Figure 2) suggest that the Southern Ocean was supplied with "young" (high O2, low nutrient, high δ13C) deep water in the latest Paleocene (approximately 60-58 Ma) and possibly during the early Eocene (approximately 57-52
Ma; Figure 2; Miller et al., 1987a; Pak et al., 1990; Katz and Miller, 1991). Oxygen isotope evidence indicates that the Southern Ocean was filled with cooler water than the Pacific beginning ca. 60 Ma (Miller et al., 1987a; Katz and Miller, 1991), supporting our contention that the deep-water source was Antarctic. However, near the Paleocene/Eocene boundary (approximately 58.5–57 Ma), the supply of Southern Ocean “young” deep water appeared to be reduced or eliminated (Figure 2). We speculate that a temporary elimination of this inferred Antarctic source triggered the benthic foraminiferal turnover. This interpretation is consistent with that of Thomas (1990) and Kennett and Stott (1991), who attributed the extinctions to a pulse of Warm Saline Deep Water (WSDW) produced in low latitudes. We concur that WSDW may have replaced deep waters that were produced near Antarctica prior to and probably subsequent to this event. This pulse of WSDW would not only have warmed the deep waters, but also would have reduced dissolved oxygen concentrations in the deep sea. Thomas (1990) reported an increase in inferred infaunal species at Site 689 and 690 which were more tolerant of low-oxygen conditions; Kaiho (1991) used benthic foraminiferal test morphology to suggest that there were reduced dissolved oxygen concentrations across this boundary in the Atlantic, Indian, and Pacific Oceans.

Available data suggest that a combination of changing physicochemical conditions triggered the extinctions. A transient, abrupt change from high to low-latitudes deep-water sources resulted in a warming and reduced oxygen in deep waters, which caused the benthic extinction event (Figure 1). These changes in deep water were probably related to global warming, possible high CO2 levels, and reduced zonal wind intensities (Owens and Rea, 1985; Miller et al., 1987a; Rea et al., 1991).

**GRADUAL MIDDLE EOCENE-EARLIEST OLIGOCENE FAUNAL CHANGES: RESPONSE TO CHANGING DEEP WATERS**

Because periodicity models predict a mass extinction at the end of the Eocene (Raup and Sepkoski, 1982) and because a
large $\delta^{18}O$ increase occurred during the earliest Oligocene (e.g., Kennett and Shackleton, 1976), benthic foraminiferal response to late middle Eocene-earliest Oligocene oceanographic changes has attracted considerable attention (Schnitker, 1979; Corliss, 1979, 1981; Miller, 1983; Tjalsma, 1983; Tjalsma and Lohmann, 1983; Clark and Wright, 1984; Snyder et al., 1984; Miller et al., 1985; Wood et al., 1985; Corliss and Keigwin, 1986; Miller and Katz, 1987b; Thomas, in press b; Bolotovskoy and Bolotovskoy, 1988; Oberhansli et al., 1991). The consensus is that benthic foraminiferal faunal changes were gradual through this interval. Generally few first and last occurrences are found associated with the Eocene/Oligocene boundary or the worldwide earliest Oligocene $\delta^{18}O$ increase.

First and last occurrences were staggered throughout the late middle Eocene to earliest Oligocene interval (Biochron P14–P18: –42–34 Ma). For example, Miller et al. (1985) showed that there was a series of first and last occurrences of taxa at Irish margin Site 549 (2515 m present depth; 2000–2500 m paleo-depth) during the late middle Eocene to earliest Oligocene. New benthic foraminiferal data from Site 703 (subantarctic eastern South Atlantic; 1976 m present depth; –1000–1500 m paleo-depth) show a similar pattern of gradual and first and last occurrences through the upper middle Eocene to lower Oligocene section (Figure 3). Boersma (1986) noted FO of two taxa and the LO of five taxa spanning the Eocene/Oligocene boundary in the southwest Pacific. Most of the first and last occurrences reported from various locations were global, although the expected timing varied locally. For example, *Nuttallides umbonifera* first appeared in the late Eocene on the Irish margin (Miller et al., 1985), and Berggren and Miller (1989) placed its global first occurrence near the beginning of the late Eocene (Figure 1). However, Katz and Miller (1991) and Thomas (in press) have recently found this taxon in upper middle Eocene sediments in the subantarctic. Other taxa which first occurred in this interval include *Astrononion pusillum*, *Bolivina testiformis*, *Bolivinopsis cubensis*, *Cassidulina havanensis*, and *Epistominel- la exigua* (e.g., Miller et al., 1985). Extinctions include *Abys- samina poagi*, *Alabamina dissonata*, *Cibicoides truncanus*, *Nuttallides truempyi*, and *Bolivinopsis spectabilis*.

A dramatic faunal abundance change occurred throughout the Atlantic (Tjalsma and Lohmann, 1983; Oberhansli et al., 1991), Caribbean (Tjalsma and Lohmann, 1983; Wood et al., 1985), Pacific (Corliss and Keigwin, 1986), and Southern Oceans (Thomas, in press; this study) during the late middle Eocene. A biofacies dominated by *Nuttallides truempyi* was replaced by a biofacies dominated by *Oridorssalis* spp., *Globobac- siulina subglobosa*, *Gyroidinoioides* spp., and *Cibicoides prae- munsulatus* (Miller, 1983; Tjalsma and Lohmann, 1983; Clark and Wright, 1984; Miller et al., 1985). The timing of this change is uncertain because of problems in late middle to earliest late Eocene biochronology. For example, the decrease in abundance of *N. truempyi* could have occurred either in the late middle Eocene (Biochron P14) or early late Eocene (Biochron P15; see figs. 10 and 11 of Miller et al., 1985 and fig. 4 of Corliss and Keigwin, 1986). Data from Site 702 show that the decrease in *N. truempyi* occurred during the late middle Eocene at this site (~42.5–40.0 Ma; Figure 4). The interbasinal correlation of this event and its correlation to the time scale remain uncertain. However, the record at Site 702 demonstrates that the decrease in abundance of *N. truempyi* was associated with a late middle Eocene $\delta^{18}O$ increase (Figure 4).

Three benthic foraminiferal $\delta^{18}O$ increases occurred during the middle Eocene to earliest Oligocene: near the early/middle Eocene boundary, in the late middle Eocene, and in the earliest Oligocene (e.g., Keigwin and Corliss, 1986; Miller et al., 1987b; Miller, in press). These $\sim 0.3$–$0.5$% $\delta^{18}O$ increases each represent some drop in deep water temperatures. For example, –2.3–2.6% in the earliest Oligocene increase has been attributed to ice volume increase, leaving 0.5–0.7% attributable to cooling (~2–4°C; Keigwin and Corliss, 1986; Miller et al., 1987b; Zachos et al., in press; Miller, in press). We note that there are similarities between the inferred coolings and the faunal record: 1) the first and last occurrences began during the late middle Eocene $\delta^{18}O$ increase and culminated at the time of the earliest Oligocene increase (Figure 3); and 2) at Site 702, *N. truempyi* decreased in abundance during the late middle Eocene $\delta^{18}O$ increase (Figure 4). However, there is no direct relationship between temperature changes and the faunal record because: 1) there were few first or last occurrences associated with the early middle Eocene $\delta^{18}O$ increase; 2) there was little change in abundance of *N. truempyi* during the early/middle Eocene $\delta^{18}O$ increase (Figure 4); and 3) first and last occurrences are scattered throughout the interval between the late middle Eocene and earliest Oligocene $\delta^{18}O$ increases (Figure 3). This suggests that temperature changes alone did not cause the relative abundance changes; we speculate that changing deep-water conditions associated with the late middle Eocene cooling caused the faunal abundance change.

In addition to a deep-water cooling, there were other significant oceanographic changes during the earliest Oligocene:

1) a pulse of northern component water, analogous to modern North Atlantic Deep Water (Miller and Tucholke,
Cibicidoides praeunludus are stratigraphically long-ranging and bathymetrically wide-ranging forms with a broad inferred environmental tolerance (Miller, 1983; Tjalsma and Lohmann, 1983; Corliss and Keigwin, 1985), and they survived the large oceanographic changes that occurred during the earliest Oligocene.

Although benthic faunas remained largely unchanged, some minor faunal abundance changes occurred during the earliest Oligocene. Corliss (1979) reported an earlier Oligocene increase in Nutallides umbonifera at Tasman Sea DSDP Site 277. Boersma (1986) noted faunal fluctuations across the Eocene/Oligocene boundary in the southwest Pacific. A large relative abundance change occurred across the Eocene/Oligocene boundary at subtropical Site 703, where a Trifarina davisii-dominant biofacies replaced an Oridorsalis-dominated biofacies (Figure 5). This faunal change began prior to the δ18O increase, and the T. davisii biofacies became established after the increase (Figure 5). We conclude that regional changes in biofacies occurred in response to the earliest Oligocene circulation changes, but that these changes were localized and relatively minor compared to the late middle Eocene faunal abundance change (Figure 4). Thus, the responses of deep-water benthic foraminifera to the Eocene-Oligocene oceanographic changes were subdued, and there was no mass extinction event as predicted by Raup and Sepkoski (1982; also see Corliss et al., 1984).

Miller and Katz (1987a) found few first or last occurrences of taxa and only minor faunal abundance changes in the abyssal western North Atlantic during the Oligocene. However, during the Oligocene, there was a series of first occurrences in benthic foraminifera in the South Atlantic (Figure 1; Tjalsma, 1983). Boersma (1985,1986) noted several important migration events in benthic foraminifera during the Oligocene. Miller (1983), Boersma (1985), and Miller and Katz (1987a) noted a distinct assemblage of Nutallides umbonifera during the “middle” Oligocene. Because the abundance of this taxon is related to carbonate availability (Bremner and Lohmann, 1982), this taxon is interpreted as reflecting increased corrosivity of deep water (Miller, 1983; Boersma, 1985). The assemblage is apparently limited geographically, and only occurs in certain regions such as the Bay of Biscay (Miller, 1983), Tasman Rise (Boersma, 1985), and intermediate depths of the South Atlantic (Miller, Tjalsma, and Katz, unpublished data). Nevertheless, the faunal abundance and migration events point toward a continually changing Oligocene deep-sea environment.

**THE MIDDLE MIOCENE DEVELOPMENT OF THE “MODERN” FAUNA**

Considerable attention has been given to early to middle Miocene deep-water benthic foraminiferal changes (Berggren, 1972; Schmittk, 1979; Woodruff and Douglas, 1981; Boersma, 1986; Thomas, 1985,1986a, b, 1989b; Woodruff, 1985; Kuhlna and Kennett, 1986; Schmittk, 1986; Miller and Katz, 1987a; Thomas and Vincent, 1988; Woodruff and Savin, 1989; Woodruff, this volume). These authors documented that large changes in taxonomic composition, relative abundances, and absolute abundances began in the late early Miocene and culminated in the middle Miocene. Deep-water benthic foraminiferal taxonomic composition has remained virtually unchanged since the middle Miocene, although species abundances have fluctuated (Miller and Katz, 1987a). In essence, the “modern” fauna developed during the middle Miocene (Berggren, 1972). Boloskiy (1980) suggested that the modern fauna evolved in the Oligocene. However, a series of early-middle Miocene first and
last occurrences introduced a new fauna that was taxonomically similar to the modern deep-sea fauna. Subsequently, there were few global first or last occurrences in the late Neogene abyssal realm (Miller and Katz, 1987a; Berggren and Miller, 1989). In addition, the late middle Miocene faunas were often strikingly similar in relative abundance composition to the present day fauna (e.g., Miller and Katz, 1987a). Thus, we concur with Berggren’s (1972) original suggestion that modern deep-sea benthic foraminiferal faunas developed in the middle Miocene.

There are still disagreements as to the precise timing of the Miocene benthic foraminiferal changes. Initial studies suggested that the faunal change began in the earliest middle Miocene (ca. 16 Ma; Biochron N8; Schnitker, 1979, 1986; Woodruff, 1985). Thomas (1985, 1986a,b, 1989) and Miller and Katz (1987a) showed that the faunal change began in the late early Miocene (Biochrons N6-N7). In the western North Atlantic, absolute abundance changes began at ca. 19 Ma (Biochron N6), while relative abundance changes and the taxonomic turnover did not begin until ca. 17-16 Ma (Biochron N7; Miller and Katz, 1987a). Thomas (1985) noted that the faunal turnover began in the late early Miocene in the Pacific (ca. 17 Ma; Biochron N7).

There are also disagreements over the cause of the Miocene benthic foraminiferal changes. Woodruff (1985) related the turnover to glacial climatic cooling associated with a major middle Miocene δ¹⁸O increase. The increase in benthic foraminiferal δ¹⁸O values has been directly tied to Cryorans C5Ar to C5Ar (14.8-12.5 Ma; Biochrons N10-N12; Miller and Fairbanks, 1985). Thus, the faunal change began prior to the δ¹⁸O increase and cannot be directly attributed to deep-water cooling (see Thomas and Vincent, 1988). Miller and Katz (1987a) speculated that a change in surface ocean productivity was the primary cause. Thomas and Vincent (1988) also favored changes in surface ocean productivity as the primary cause, coupled with increased variability of deep-water corrosiveness. Woodruff and Savin (1989) favored changes in water masses as the primary trigger for the faunal changes. We believe that the faunal changes were related in some manner to both changing deep-water sources and/or surface productivity changes that may be linked. For example, Miller and Katz (1987a) and Thomas (1987) noted that the faunal changes began at about 19 Ma, the same time as a pulse of Northern Component Water (Miller and Fairbanks, 1985; Wright, Miller, and Fairbanks, in press). However, we believe that it is premature to identify a cause of the middle Miocene turnover because the first-order interpretation of deep-water patterns for the Miocene is still debated (cf., Miller and Fairbanks, 1985 with Woodruff and Savin, 1989).

SUMMARY

We reviewed the nature and cause of three Cenozoic deep-water benthic foraminiferal turnovers:

1. The exact timing of the latest Paleocene turnover is still uncertain, although it appears clear that the "crisis" occurred very rapidly during the latest Paleocene. New data (Kennett and Stott, 1991) suggest that the cause of this crisis was abrupt, transient change in deep-water sources coupled with an abrupt warming. The extinctions may have resulted from an influx of warm saline deep water (Thomas, 1990; Kennett and Stott, 1990) that temporarily replaced Antarctic sources of deep water (Miller et al., 1987a; Katz and Miller, 1991).

2. The timing of the late middle Eocene decrease in abundance of *Nostilidites truempyi* is poorly correlated to the geomagnetic polarity time scale. We directly link this decrease in abundance to an increase in benthic foraminiferal δ¹⁸O values, and attribute it to changing deep-water conditions associated with a late middle Eocene cooling. Beginning with this abundance change, deep-water benthic foraminifera experienced a series of first and last occurrences that may have been a response to changing deep-water conditions.

3. A taxonomic turnover and faunal abundance changes also occurred beginning in the late early Miocene (ca. 19 Ma; Biochron N6). These changes culminated in the middle Miocene (ca. 12.5 Ma; Biochron N12) with the development of the modern fauna. The cause of this change remains debatable, although most authors suggest changing deep-water sources and/or surface ocean productivity.

ACKNOWLEDGMENTS

Our studies of deep-water benthic foraminifera have benefited from interactions with many colleagues, including J. Aubert, R.C. Tjalsma, A. Boersma, B. Coriliss, D. Schnitker, E. Thomas, F.P.C.M. van Morkhoven, and F. Woodruff. Discussions with E. Thomas on the Paleocene/Eocene turnover have been particularly useful. We thank A. Boersma, D.K. Pak, and E. Thomas for reviews. This work was supported by NSF grants OCE88-17563, OCE90-19569, and a consortium of oil companies (BP, Chevron, Marathon, Texaco, and Unocal). Samples were supplied by the ODP. This is L-DGO contribution number 4936 and Woods Hole Oceanographic Institution contribution number 7692.

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