BIOCHRONOLOGY AND PALEOCLIMATIC IMPLICATIONS OF MIDDLE EOCENE TO OLIGOCENE PLANKTIC FORAMINIFERAL FAUNAS

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Abstract

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Planktic foraminiferal assemblages have been analyzed quantitatively in six DSDP sites in the Atlantic (Site 363), Pacific (Sites 292, 77B, 277), and Indian Ocean (Sites 219, 253) in order to determine the nature of the faunal turnover during Middle Eocene to Oligocene time. Biostratigraphic ranges of taxa and abundance distributions of dominant species are presented and illustrate striking similarities in faunal assemblages of low latitude regions in the Atlantic, Pacific and Indian oceans. A high resolution biochronology, based on dominant faunal characteristics and 55 datum events, permits correlation between all three oceans with a high degree of precision. Population studies provide a view of the global impact of the paleoclimatic and paleoceanographic changes occurring during Middle Eocene to Oligocene time.

Planktic foraminiferal assemblage changes indicate a general cooling trend between Middle Eocene to Oligocene time, consistent with previously published oxygen isotope data. Major faunal changes, indicating cooling episodes, occur, however, at discrete intervals: in the Middle Eocene 44-43 Ma (P13), the Middle/Late Eocene boundary 41-40 Ma (P14/P15), the Late Eocene 39-38 Ma (P15/P16), the Eocene/Oligocene boundary 37-36 Ma (P18), and the Late Oligocene 31-29 Ma (P20/P21). With the exception of the E/O boundary, faunal changes occur abruptly during short stratigraphic intervals, and are characterized by major species extinctions and first appearances. The Eocene/Oligocene boundary cooling is marked primarily by increasing abundances of cool water species. This suggests that the E/O boundary cooling, which marks a major event in the oxygen isotope record affected planktic faunas less than during other cooling episodes. Planktic foraminiferal faunas indicate that the E/O boundary event is part of a continued cooling trend which began during the Middle Eocene.

Two hiatus intervals are recognized in low and high latitude sections at the Middle/Late Eocene boundary and in the Late Eocene (P15/P16). These hiatuses suggest that vigorous bottom water circulation began developing in the Middle Eocene, consistent with the onset of the faunal cooling trend, and well before the development of the psychrosphere at the E/O boundary.

Introduction

A multitude of geological events with world wide impact occurred during the Middle Eocene to Early Oligocene. All evidence points to this time interval as the climatically most unstable period of the Tertiary ultimately leading to the modern climatic and oceanic circulation system. Major biological events include successive extinctions of tropical marine faunas and floras and their replacement by cooler subtropical and temperate elements (Haq and Lohmann, 1976; Haq et al., 1977; Sancetta, 1979; Berggren, 1978; Kennett, 1977, 1978; McGowran, 1978). These biogeographic changes were accompanied by a gradual decrease in $\delta^{18}O$ inferred paleotemperatures of both surface and bottom waters, but with

a significantly larger temperature drop in bottom waters near the Eocene/Oligocene boundary (Douglas and Savin, 1975; Shackleton and Kennett, 1975; Keigwin, 1980; Miller and Curry, 1982). Moreover, a global deepening in the calcite compensation depth (CCD) of nearly 1000 m (from 3000 to 4000 m) occurred during this time (Berger, 1972; Van Andel et al., 1975) associated with lowering of the global sea level (Vail et al., 1977; Vail and Hardenbol, 1979), and major hiatuses (Van Andel et al., 1975; Moore et al., 1978; McGowran, 1978; Kennett et al., 1972).

Corresponding changes also occurred in terrestrial faunas and floras, as for instance, the major faunal change described as the "Grand Coupure" which gave rise to the evolution of larger land mammals (Stehlin, 1909; Brunet, 1979; Cavalier, 1979). Major floral changes include the replacement of angiosperms by gymnosperms (Van Couvering et al., 1981) and replacement of evergreen forests by deciduous forests in middle to high latitudes indicating a general cooling trend at this time (Wolfe, 1978). These phenomena collectively are known as the "Terminal Eocene Event" (Van Couvering et al., 1981) which denotes the climax near the E/0 boundary of a succession of events, all of which are in some way connected to an irreversible global cooling trend.

This cooling trend is related in part to the separation of Australia from Antarctica which initiated the Antarctic glacial development in the mid Tertiary (Kennett et al., 1975). Cooling accelerated in the Middle to Late Eocene partly as a result of a shallow connection (200–300 m) which linked the southwest Pacific with the southeast Indian Ocean across the South Tasman Rise (Kennett et al., 1975) creating a prototype of the Circum-Antarctic Current. This development resulted in increased thermal isolation of the Antarctic (Hamilton, 1968; Donn and Shaw, 1977) which led to increased global cooling (Kennett, 1977, 1978).

The major oxygen-isotope inferred high-latitude temperature drop occurred near the E/0 boundary and substantially lowered

bottom water temperatures. This drop in oceanic bottom-water temperatures has been interpreted to mark the development of the psychrosphere (two layer ocean with lower cold layer temperatures <10°C) believed to reflect a decrease in Antarctic surface water temperatures to near freezing and the development of vigorous bottom water circulation (Shackleton and Kennett, 1975), or the production of the first extensive sea ice near Antarctica (Matthews and Poore, 1980). Alternatively, the development of the psychrosphere may be viewed as part of the vigorous bottom-water circulation which began to develop in the Middle to Late Eocene reflecting a gradual climatic cooling (Corliss, 1981). Extensive deep-sea erosion recognized in Middle to Late Eocene sequences in earlier studies (Kennett et al., 1972; Van Andel et al., 1975; Moore et al., 1978; McGowran, 1978) and this report, also imply intensified bottom-water circulation started as early as the Middle Eocene.

The paleoclimatic history of the oceans can best be studied in the changing faunal and floral assemblages in response to climatic and oceanographic changes. This study attempts to reconstruct the paleoclimatic and biostratigraphic history of the Middle Eocene through Oligocene based on quantitative analyses of planktic foraminifers in the Atlantic, Pacific and Indian Oceans using the best preserved sections available (DSDP Sites 363, 292, 77B, 277, 219, 253, Fig. 1). Major results of this report include: (1) a gradual cooling trend occurs in planktic foraminiferal assemblages between Middle
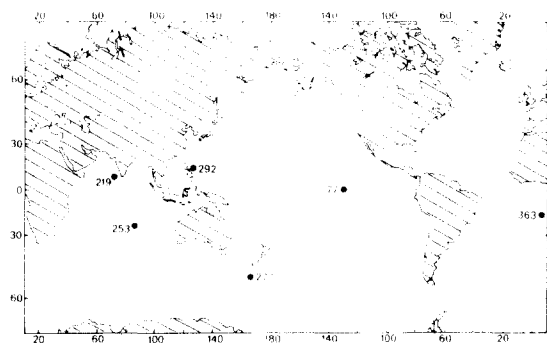

Fig. 1. Location map of DSDP Sites examined.

Eocene to Oligocene time; (2) major faunal assemblage changes occur at discrete intervals generally correlating with isotopic cooling; (3) there are no catastrophic extinctions at the E/O boundary; (4) widespread hiatus distribution in Middle and Late Eocene deep-sea sequences indicate vigorous bottom water circulation started well before the development of the psychrosphere at the E/O boundary.

Methods

Quantitative population distributions of planktic foraminifers have been analyzed in Middle Eocene to Oligocene deep-sea sequences and provide the basis for paleoclimatic and biochronologic interpretations. Paleoclimatic conditions are interpreted from fluctuations in the abundance of temperature sensitive species and their changing biogeographic distributions. Biochronologic resolution has been improved through the combined use of planktic zones and datum levels. In addition, the quantitative faunal record permits more accurate correlation between oceans and aids in recognizing short hiatuses based on rapid abundance changes of individual species. Furthermore, analyses of low latitude deep-sea sections in the Atlantic, Pacific and Indian oceans demonstrate that the same zonal system can be used in each ocean. This high resolution biochronology and ability to correlate between oceans thus permits more accurate timing of paleocirculation changes and paleoclimatic events. Methods of sample preparation and quantitative population studies have been described in Keller (1980).

Biochronology

Numerous zonal systems have been published for the Paleogene, but the most commonly used are those of Bolli (1957a, b, 1966, 1972), Blow (1969) and Stainforth et al. (1975). In this study Bolli's zonal scheme has been found to be easily applicable in the Atlantic, Pacific and Indian oceans. Therefore, Bolli's zonation is used here with

the exception of the *Cassigerinella chipolensis—Pseudohastigerina micra* Zone which is modified to *P. micra* Zone because *C. chipolensis* is solution-susceptible and only sporadically present. Similarly, Berggren (1972) and Stainforth et al. (1975) found Bolli's zonal scheme with minor revisions most applicable. Strict definitions of Bolli's zones as well as numerous other zonal systems are discussed in Berggren (1972), Stainforth et al. (1975), and Hardenbol and Berggren (1978).

Blow's (1969) P-zones, while convenient for the short letter-number designation, are often difficult to apply because of the use of transitional and solution-susceptible species. Difficult to identify transitional species, such as *Globigerina pseudoampliapertura* and *G. tapuriensis*, and solution-prone species introduce additional uncertainties in placing zonal boundaries.

In any detailed biostratigraphic study applying a zonal scheme is a first approximation of the completeness of a sedimentary record. Absence of zonal index markers may indicate either preferential dissolution of species, absence due to paleoecologic factors, or a hiatus caused by erosion or nondeposition. Presence of zonal index markers indicates that the record *may* be complete. Even if all zonal markers are present, however, a short hiatus can be present within a zone which may thus go undetected. Such short breaks in the geologic record may be caused by nondeposition due to increased carbonate dissolution and can be identified through the use of multiple datum events (first and last appearances of species) within zones. For instance, the first appearance of *Globorotalia cerroazulensis cunialensis* occurs near the middle of the *G. cerroazulensis* s.l. Zone (Toumarkine and Bolli, 1975). In the deep-sea sections analyzed here the first occurrence of this species falls near the base of this zone indicating that the lower part of the *G. cerroazulensis* Zone is removed by a hiatus.

The most valuable tool in recognizing short hiatuses, however, is the quantitative abundance record. This can be most aptly
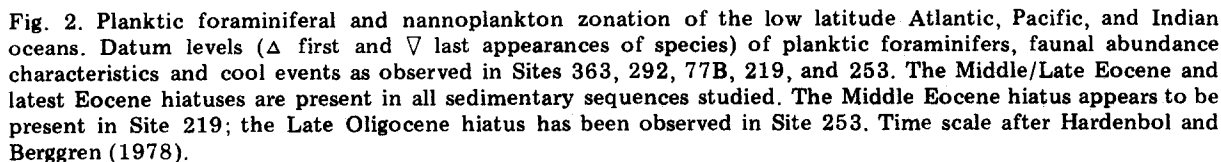
Age (m.y)	Epochs	Nannoplankton		Planktic Foraminifera		Datum Events	Faunal Characteristics	Hiatus	Cold Events
		Martini 1971	Bukry 1973	Blow 1969	Bolli 1957a,b 1966 1972	▲ INDEX SPECIES △ FIRST APPEARANCE FA ▽ LAST APPEARANCE LA • FA LA VARIABLE			

Fig. 2 table (partially legible):

Age	Epochs	Nanno (Martini)	Nanno (Bukry)	Blow	Bolli
30	OLIGOCENE	NP23	Sphenolithus distentus	P 21	G. opima
32	OLIGOCENE		Sphenolithus predistentus	P 20	Globigerina ampliapertura
		NP22	R. hillae SZ	P 19 & P 18	Pseudohastigerina micra
37	LATE EOCENE	NP21	C formosus & C.subdistichus SZ	P 17	Globorotalia cerroazulensis s.l.
38	LATE EOCENE	NP20 & NP19	I. recurvus SZ	P 16	
				P 15	Globigerinatheka semiinvoluta
40		NP18	C oamaruensis	P 14	Truncorotaloides rohri
	MIDDLE EOCENE	NP17	D saipanensis SZ		
44			D. bifax SZ	P 13	Orbulinoides beckmanni
		NP16 & NP15	C. staurion SZ	P 12	Globorotalia lehneri
			C. gigas SZ	P 11	Globigerinatheka subconglobata

Fig. 2. Planktic foraminiferal and nannoplankton zonation of the low latitude Atlantic, Pacific, and Indian oceans. Datum levels (△ first and ▽ last appearances of species) of planktic foraminifers, faunal abundance characteristics and cool events as observed in Sites 363, 292, 77B, 219, and 253. The Middle/Late Eocene and latest Eocene hiatuses are present in all sedimentary sequences studied. The Middle Eocene hiatus appears to be present in Site 219; the Late Oligocene hiatus has been observed in Site 253. Time scale after Hardenbol and Berggren (1978).

illustrated by *Globigerinatheka semiinvoluta* whose range characterizes the zone by the same name. The base and the top of this zone are frequently marked by hiatuses. These hiatuses are generally not recognized, unless the preceding or succeeding zones are also removed. In a quantitative record, however, these hiatuses can easily be detected by the abrupt abundance changes of species. For instance, at Site 363 *Truncorotaloides rohri* comprises 40% of the fauna in one sample, and *G. semiinvoluta* is absent. In the next sample, 20 cm above this interval, *T. rohri* is absent whereas *G. semiinvoluta* comprises 10%. Similarly, other species appear and disappear with equally sudden abundance changes.

The quantitative record is also useful in correlating sites, even from one ocean basin to another, provided that similar oceanic conditions prevailed over the region to be correlated. Detailed analyses of DSDP sites in the Atlantic (Site 363), Pacific (Sites 292, 77, 277), and Indian Ocean (Sites 219, 253) has indeed revealed that the faunal succession as well as percent abundance fluctuations of dominant species in low latitude regions of each ocean are very similar. Moreover, abundance peaks of individual species occur at the same time intervals presumably in response to changing paleoceanographic conditions. These abundance peaks are invaluable both as aid in stratigraphic correlation and in paleoclimatic interpretation.

Correlation of these faunal characteristics, datum events and hiatuses to the zonal schemes of Bolli and Blow, as observed in this study, are illustrated in Fig. 2 along with the coccolith zonation of Bukry (1973) and Martini (1971). Bukry's nannofossil zonation has been used here. In the deep-sea sites analyzed in this study, the relative position of the nannofossil zone boundaries to the planktic foraminiferal zones has been found to be most consistent when Bukry's site report data were used. This illustrates the uncertainties introduced when correlating biostratigraphic data from several workers. Nevertheless, the sequence of datum events

agrees well with the nannofossil and planktic foraminiferal events recently discussed by Beckmann et al. (1981) and Van Couvering et al. (1981). In the following section the main biostratigraphic and faunal characteristics of Sites 363, 219, 292, 77B, 253 and 277 are discussed. Species identifications were aided by the works and illustrations of Bolli (1957a, b) Loeblich and Tappan (1957), Toumarkine and Bolli (1970, 1975), Stainforth et al. (1975) and Blow (1969), as well as by the authors of the Initial Reports of the Deep Sea Drilling Project as discussed below.

Two low latitude deep-sea sequences, Sites 363 (South Atlantic) and Site 219 (Indian Ocean) provide the most complete Middle Eocene through Early Oligocene record to date. Two equatorial Pacific sequences, Sites 292 and 77B are notable for their excellent representations of latest Eocene through Oligocene records. The high southern paleolatitude Indian Ocean Site 253 and South Pacific Site 277 are notable for their differences, with Site 253 showing greater faunal affinities with low latitude faunas whereas Site 277 has subantarctic to temperate assemblages.

Biostratigraphy

Site 363

South Atlantic DSDP Site 363 is located on the Walvis Ridge at 2248 m water depth. Backtracking indicates that Site 363 was at 2000—2100 m depth during Middle Eocene to Early Oligocene time (Berger and Winter, 1974). The paleolatitude of this Site has changed little (from 21°S to 19°38'S) since the Cretaceous (Ryan et al., 1978).

Nannoplankton at DSDP Site 363 were initially studied by Proto Decima et al. (1978) and Bukry (1978). Bukry's report was used here, except for the addition of the *Chiasmolithus oamaruensis* Zone of Proto Decima et al. (1978) which Bukry did not recognize due to wider sample spacing. The planktic foraminiferal biostratigraphy was studied by Toumarkine (1978) and only
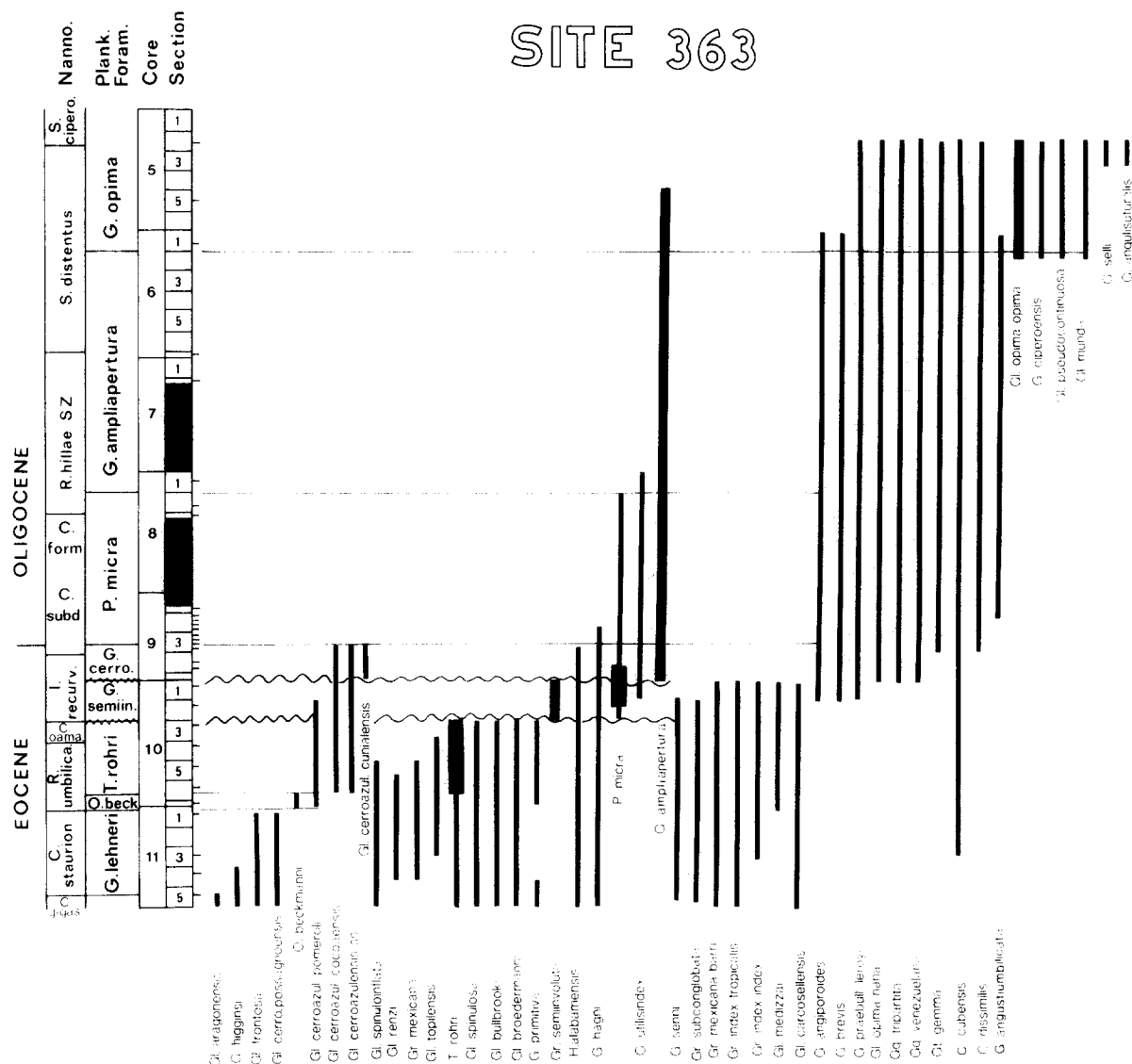
Fig. 3. Ranges of selected planktic foraminiferal species of Site 363, South Atlantic. Thick lines mark relative abundance of index species; wavy lines mark hiatuses; tick marks in core-section indicate sample locations. Nannoplankton zonation after Bukry (1978) and Proto Decima et al. (1978).

minor changes in the position of zonal boundaries, due to closer sample spacing, are reported here. No hiatuses were recognized by these workers.

The zonation, ranges of diagnostic species, and abundance distributions of dominant species are illustrated in Figs. 3 and 4. Each planktic foraminiferal zone can be charac-

terized by the abundance fluctuations of dominant species, and this sequence of faunal succession greatly improves biostratigraphic correlation and biochronologic resolution. Thus, the Middle Eocene *Globorotalia lehneri* Zone is characterized by declining abundances of *Globorotalia broedermanni* and *G. bullbrooki* and increasing *G.*
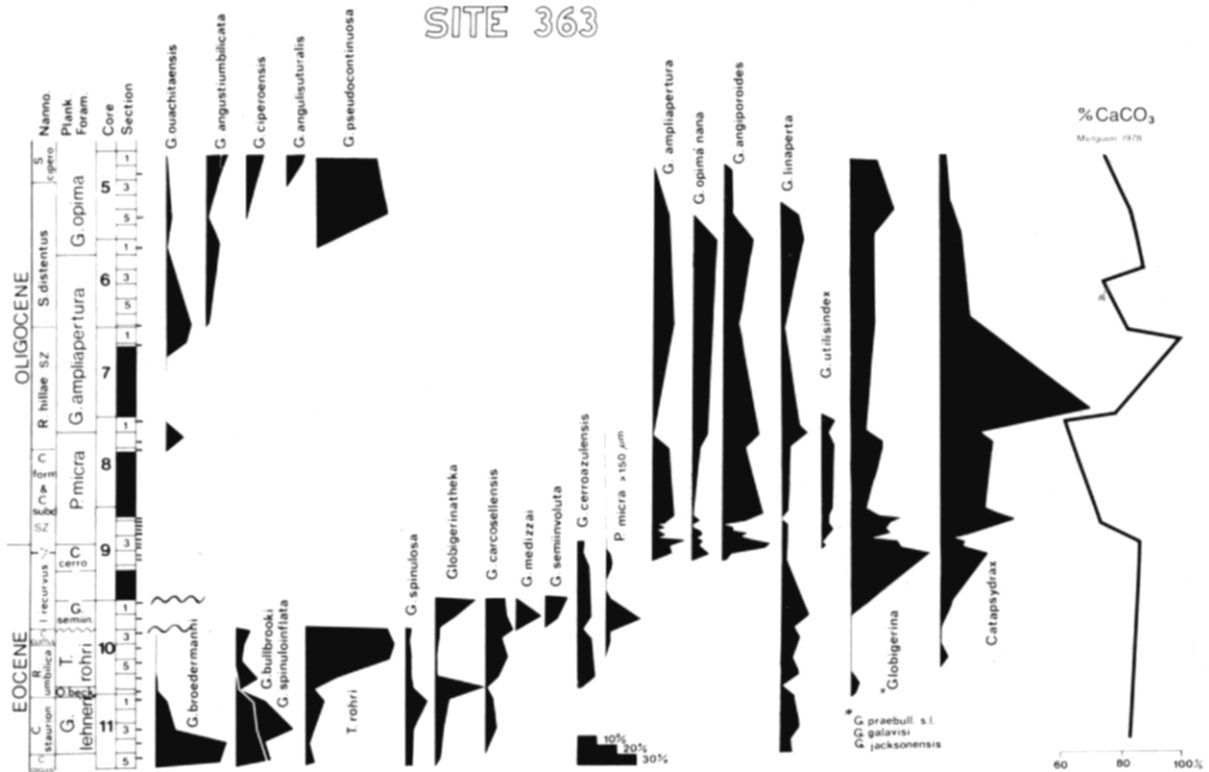
Fig. 4. Abundance distributions of dominant planktic foraminiferal species of Site 363, South Atlantic. Note the abrupt faunal changes in the Middle/Late and latest Eocene mark hiatuses (wavy lines). Nannoplankton zonation after Bukry (1978) and Proto Decima et al. (1978). Percent carbonate data from Melguen (1978). Tick marks in core-section indicate sample locations.

spinulosa and Truncorotaloides rohri near the top of the zone. The short Orbulinoides beckmanni Zone is often difficult to identify and is found here only in one sample (10cc, Fig. 3). The faunal composition of this interval is quite different from samples above and below and is dominated by Globigerinatheka (25%), Globigerina eocaena (22%) and G. senni (10%). The latest Middle Eocene Truncorotaloides rohri Zone (P14) is dominated by T. rohri (45%) which disappears at the hiatus that marks the middle/late Eocene boundary (Fig. 3). Spinose species disappear at this interval with the exception of occasional rare forms of T. rohri which may be reworked.

A second hiatus occurs in the latest Eocene at the G. semiinvoluta/G. cerroazulensis s.l. Zone boundary. This hiatus is marked by the simultaneous disappearance of Globorotaloides carcosellensis, Globigerina medizzai and species of Globigerinatheka. Incomplete recovery of Core 9 makes it difficult to determine the extent of this latest Eocene hiatus at Site 363. The very short presence of G. semiinvoluta Zone (Core 10, sections 1 and 2) and the presence of Globorotalia cerroazulensis cunialensis a short distance above this interval (Core 9, section 4) suggests that, similar to other deep-sea sequences, the lower half of Globorotalia cerroazulensis Zone (P16, Fig. 2) and most of the G. semiinvoluta Zone is missing. The remaining G. semiinvoluta Zone is characterized by abundance peaks in G. semiinvoluta (small specimens typical of the Mediterranean region, Toumarkine, 1978), as well as other species of Globigerinatheka, Globo-

rotaloides carcosellensis, *G. medizzai*, and *Pseudohastigerina micra* (20%, >150 μm) which appear for the first time in abundance in the size fraction greater than 150 μm (Fig. 3). The occurrence of abundant *P. micra* in the >150 μm size fraction in the *Globigerinatheka semiinvoluta* and *Globorotalia cerroazulensis* Zones provides an important biostratigraphic marker for Late Eocene to earliest Oligocene faunas. In Middle Eocene and late Early Oligocene faunas *P. micra* is commonly present in the <150 μm size fraction only.

The Eocene/Oligocene boundary in planktic foraminifers is defined by the extinction of *Hantkenina* and the *G. cerrozaulensis* group which at Site 363 occurs in Core 9-3 (107 cm) and correlates to the top of the *Ismolithus recurvus* nannofossil Subzone (Figs. 3, 4). A similar relationship between these two planktic groups at the E/O boundary has also been observed in European sections (Van Couvering et al., 1981). The most notable faunal changes in the upper part of the *G. cerrozaulensis* s.l. Zone (P17) are the increasing abundances of seemingly cooler water species *Globigerina ampliapertura*, *G. angiporoides*, *G. praebulloides*, *G. galavisi*, and *Catapsydrax* (Fig. 3). This trend continues into the Early Oligocene *P. micra* Zone. Oxygen isotope data indicate a major cooling trend at this time in oceanic bottom-water temperatures (Keigwin, 1980). Oxygen isotope data of DSDP Site 363 (L.D. Keigwin, unpublished data) indicate that the maximum paleotemperature drop occurs in Core 8 of the *P. micra* Zone, coincident with the decrease in percent $CaCo_3$ (Melguen, 1978) (Fig. 4). Unfortunately, the lower part of Core 8 is missing and the major isotopic cooling can not be placed accurately.

The lower part of the *Globigerina ampliapertura* Zone (*Reticulofenestra hillae* nannofossil Subzone, Cores, 8, 7) contains Braarudosphaera blooms coinciding with an increase in percent $CaCo_3$ and strongly reduced numbers of foraminifers presumably due to dilution. The marked increase in Catapsydrax evident at this interval (Fig. 3) may be due to dissolution in some samples.

Another notable change during the *G. ampliaperture* Zone in Site 363 is the increase in *Globigerina ouachitaensis*.

A marked assemblage change occurs in the *Globorotalia opima* Zone. At this time the latest Eocene to Early Oligocene fauna consisting of *G. linaperta*, *G. angiporoides*, *G. utilisindex*, *G. ampliapertura*, and *P. micra* is replaced by a faunal assemblage characterized by abundant *Globorotalia pseudocontinuosa*, *G. opima*, and less common *Globigerina ciperoensis* and *G. angustiumbilicata*. Foraminiferal faunas at Site 363 thus indicate a successive replacement of diverse Middle Eocene faunal assemblages by Late Eocene to Oligocene faunal assemblages of low species diversity and simple species morphology.

Site 219

Site 219 is located in the Indian Ocean on the Laccadive—Chagos Ridge (9°01.75'N, 72°52.67'E) at 1764 m water depth. During Paleocene and Eocene time an open shallow water neritic environment existed at about 100 m depth, at a paleolatitude of about 3°S (Sclater et al., 1977).

Nannoplankton were initially studied by Boudreaux (1974) and Bukry (1974) and planktic foraminifers by Fleischer (1974). The nannoplankton reports differ considerably in their assignment of zonal boundaries. When compared to the planktic foraminiferal zones of this report and to other sites, the zonal assignment of Boudreaux has been found to be more consistent for the Late Eocene and that of Bukry for the Middle Eocene. Fleischer's biostratigraphic interpretation differs from this report primarily in the Middle Eocene and the placement of the E/O boundary. These differences are most likely due to reworking and downcore contamination found throughout this sequence. These problems can be more easily avoided in quantitative faunal analysis.

Site 219 is essentially similar, but more complete, than the South Atlantic Site 363 (Figs. 5, 6). The Middle Eocene *Globierinatheka subconglobata* Zone (P11) is repre-
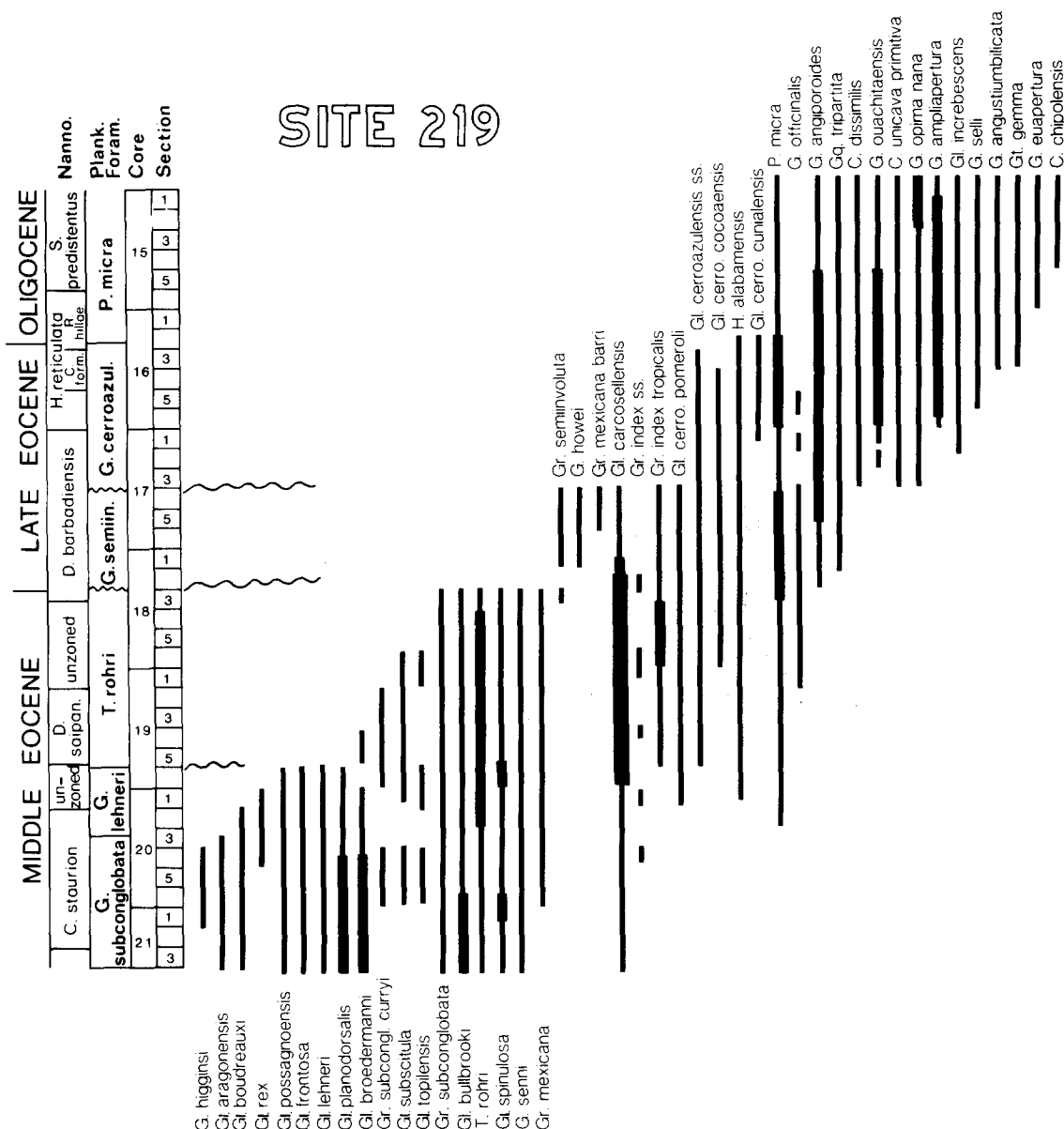
Fig. 5. Ranges of selected planktic foraminiferal species of Site 219, Indian Ocean. Thick lines mark relative abundance of index species; wavy lines mark hiatuses.

sented by abundant *Globorotalia broedermanni* (30%) and common *G. planodorsalis*, a new species described by Fleischer (1974). The top of this Zone has been placed at the last occurrence of *Globorotalia aragonensis* (Fig. 5). In Site 363 the top of this Zone is present in Core 11-5 where *G. broedermanni* is also dominant. The *Globorotalia lehneri* Zone (P12) is recognized between the top of *G. aragonensis* and the top of *G. plano-*

dorsalis, G. lehneri, G. cerroazulensis possagnoensis and *G. frontosa* (Core 19-6, Fig. 5). The latter two species also characterize this interval in Site 363. Abundance changes in the *G. lehneri* Zone are similar in the two sites, depicting a marked decline in *G. broedermanni* and *G. planodorsalis* and increase in *T. rohri* (Fig. 4). The *Orbulinoides beckmanni* Zone was not observed.
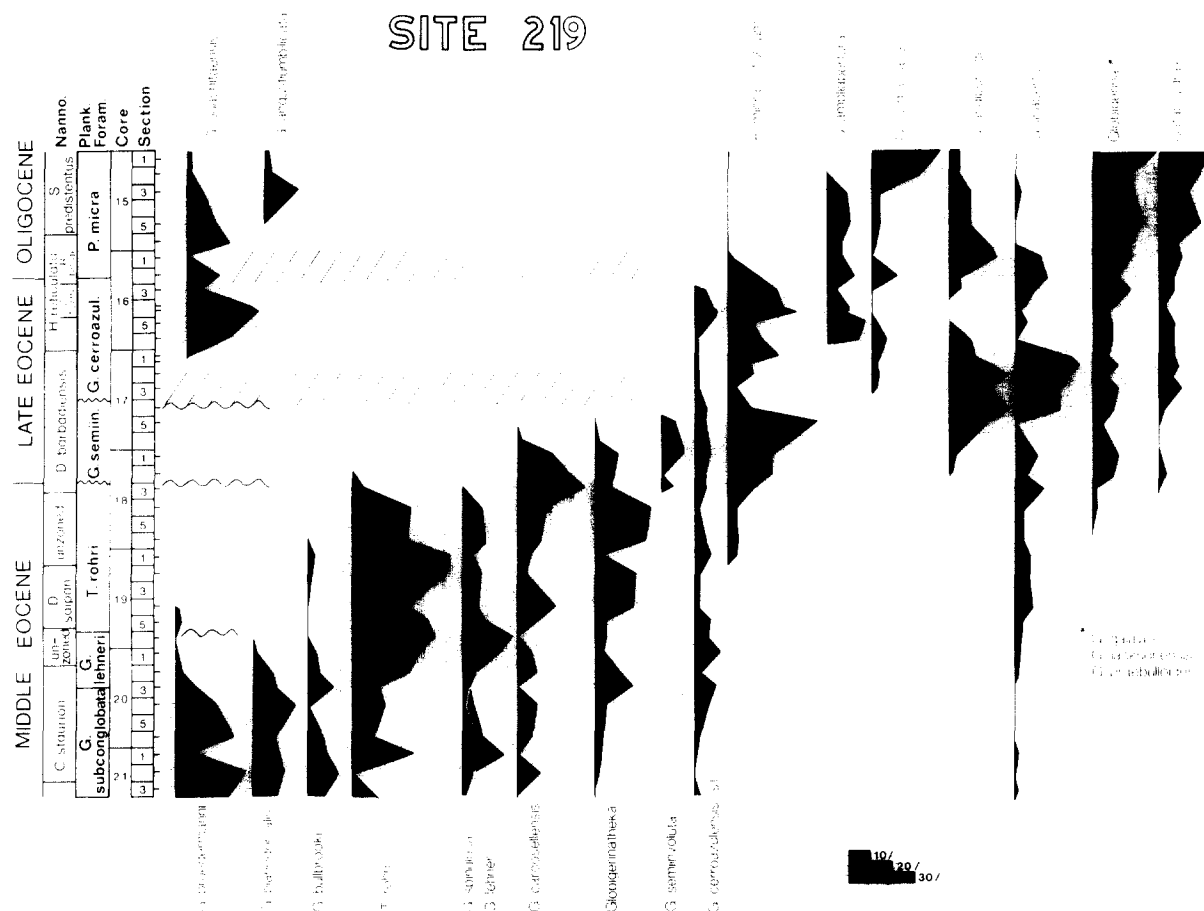
Fig. 6. Abundance distributions of dominant planktic foraminiferal species of Site 219, Indian Ocean. Wavy lines mark hiatuses; diagonal lines mark intervals of increased carbonate dissolution; tick marks on core section indicate sample locations. Nannoplankton zonation after Boudreaux (1974) and Bukry (1974).

Fleischer (1974) did not recognize the *G. lehneri* Zone and therefore argued for the existence of a significant hiatus in Core 19-6. There appears to be only a short hiatus present at this interval, as suggested by the absence of the *O. beckmanni* Zone (P13). Similarly, Bukry (1974) did not recognize the *Discoaster bifax* Subzone (of the *R. umbilica* Zone) which correlates to the *O. beckmanni* Zone and lower part of the *T. rohri* Zone and thus provides additional support for the presence of a short hiatus.

The *Truncorotaloides rohri* Zone in Sites 219 and 363 is characterized by abundant *T. rohri* (40%) and variable abundances of *Globigerinatheka* (Figs. 4, 6). As at Site 363, a hiatus is present in Site 219 at the top of this zone at the Middle/Late Eocene boundary as indicated by the simultaneous dis-

appearance of *Globigerina subconglobata, G. mexicana, Globigerina senni, Globorotalia bullbrooki, G. spinulosa* and *T. rohri* (Figs. 5, 6).

As at Site 363 the *G. semiinvoluta* Zone is characterized by common *G. semiinvoluta* (small specimens similar to Site 363), first abundant *Pseudohastigerina micra* >150 μm (20—40%) and increasing abundances in *Globigerina praebulloides, G. galavisi,* and *G. angiporoides.* The latest Eocene hiatus is also present here between the *G. semiinvoluta* and *G. cerroazulensis* Zones (Fig. 5) (Bukry, 1974; Boudreaux, 1974; Fleischer, 1974). Similar to Site 363, the upper limit of this hiatus falls shortly below the first appearance of *G. cerroazulensis cunialensis* suggesting that most of Blow's (1969) Zone P16 (lower part of *G. cerroazulensis* Zone)

is missing (Figs, 5. 6). The remaining lower part of the *G. cerroazulensis* Zone is dominated by *G. angiporoides* and *G. linaperta* (25—30%) whereas the upper part (P17) is marked by the second abundance peak in *P. micra* >150 μm (25—30%), first common *G. ampliapertura*, and first abundant *G. ouachitaensis*.

The extinctions of the *G. cerroazulensis* group and *Hantkenina alabamensis*, which characterize the foraminiferal E/O boundary, were observed here in Core 16-2 (Fig. 5),

rather than at the base of Core 16 as reported by Fleischer (1974). This places the foraminiferal E/O boundary higher in the *Helicosphaera reticulata* nannoplankton Zone than generally observed. This discrepancy, however, may be due, in part, to the differing interpretations of nannoplankton zonation at this interval (Bukry, 1974; Boudreaux, 1974).

Site 253

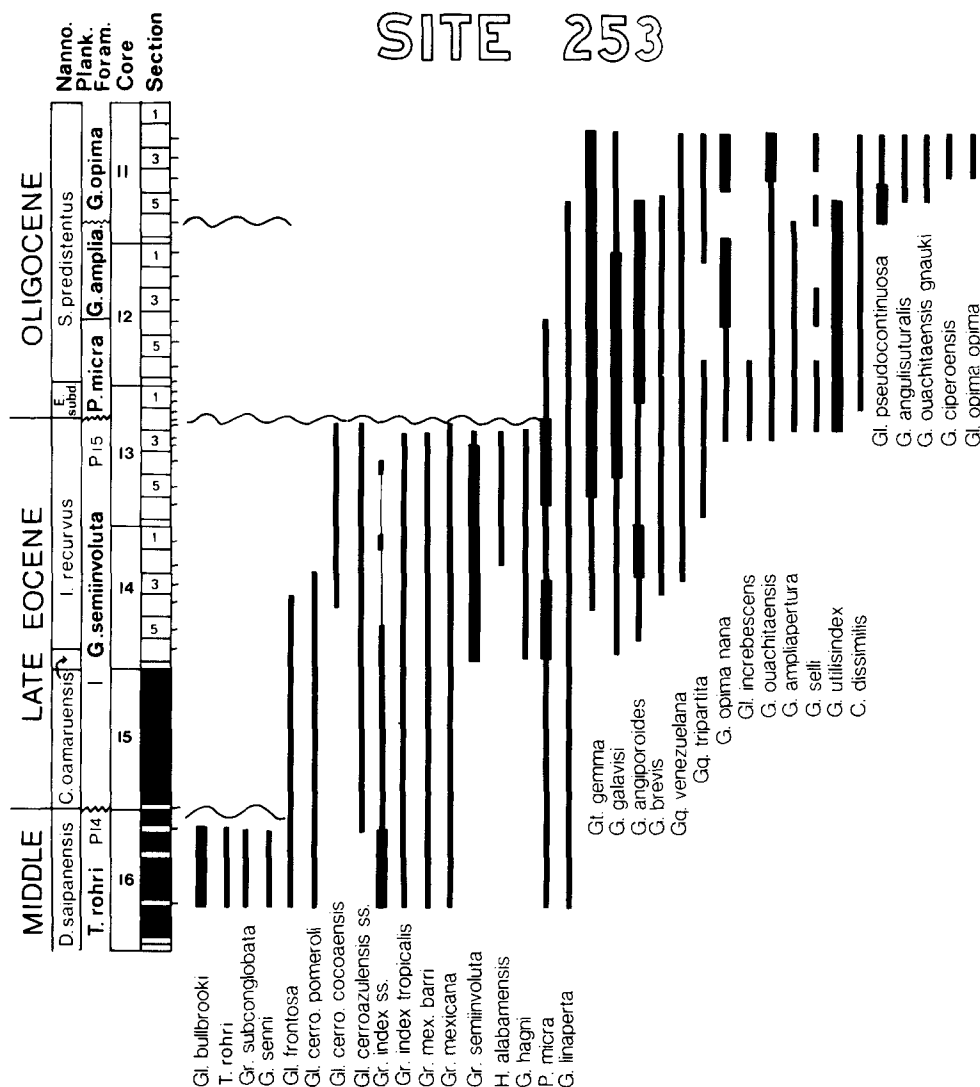Site 253 is located in the Indian Ocean on

Fig. 7. Ranges of selected planktic foraminiferal species of Site 253, southern Indian Ocean. Thick lines mark relative abundance of index species; wavy lines mark hiatuses; tick marks on core-section indicate sample locations. Nannoplankton zonation after Thierstein (1974).

Fig. 8. Abundance distributions of dominant planktic foraminiferal species of Site 253, southern Indian Ocean. Wavy lines mark hiatuses; tick marks on core section indicate sample locations. Nannoplankton zonation after Thierstein (1974).

the Ninetyeast Ridge (24°52.6'S 87°21.9'E) at a water depth of 1962 m. Backtracking indicates that during Middle Eocene time Site 253 was at 350 m depth and gradually subsided to 900 m depth by Middle Oligocene time (Sclater et al., 1977; Kidd and Davies, 1978). The paleolatitude is estimated between 43°S and 51°S (McKenzie and Sclater, 1971; Pierce et al., 1974).

Earlier biostratigraphic reports of this site include nannoplankton by Thierstein (1974) whose zonation is reproduced here, and Oligocene foraminifers by Boltovskoy (1974). No report on Eocene foraminifers has been published.

Ranges and abundance distributions of dominant planktic foraminifers indicate that the geologic record at Site 253 is similar to that of Sites 363 and 219. Species diversity, however, is lower reflecting the proximity to cool subantarctic waters during Eocene

time. Core recovery is poor in Cores 16 and 15; however, the sudden disappearance of G. bullbrooki (40%), T. rohri, G. senni and G. subconglobata suggest a hiatus may be present also between the Middle and Late Eocene (Figs. 7, 8). The G. semiinvoluta Zone is dominated by Globigerinatheka, Globigerina linaperta, and Catapsydrax. Contrary to the record at Sites 219 and 363, Pseudohastigerina micra >150 μm is not common, although smaller specimens are abundant as at Site 277.

The latest Eocene hiatus juxtaposes the Early Oligocene P. micra Zone over the G. semiinvoluta Zone, thus removing the interval represented by G. cerroazulensis Zone. This may indicate that the subsurface currents which created the hiatus were stronger and predominated longer in the high latitude southern Indian Ocean than in low latitudes, or that the increased erosion

was due to the shallower depth of Site 253.

Cooler high latitude conditions are indicated in the *P. micra* and *G. ampliapertura* Zones by the increasing abundance in *Globigerina angiporoides*, *G. utilisindex* and *Globorotalia opima nana* (Fig. 8). An abrupt faunal change indicating the presence of a hiatus occurs between Cores 11-6 and 11-5 (Figs. 7, 8). This interval is marked by the sudden appearance of abundant *Globorotalia pseudocontinuosa* (42%), first appearance of *Globigerina ouachitaensis gnauki*, *G. angulisturalis*, *G. ciperoensis* and *Globorotalia opima opima* and the disappearance of *Globigerina linaperta*, *G. angiporoides*, *G. utilisindex* and *G. ampliapertura* (Fig. 7). Faunal correlations between Sites 363, 292, and 253 suggest that the lower part of the *G. opima* Zone and much of the *G. ampliapertura* Zone is missing. This hiatus correlates to the major drop in the eustatic sea level recognized by Vail and Hardenbol (1979) in the lower part of the *G. opima* Zone.

Site 292

Site 292 is situated in the western Pacific Ocean on the Benham Rise (15°49.11′N 124°39.05′E) at 2943 m water depth. During Late Eocene time deposition occurred near the Equator at about 500 m depth with gradual subsidence to 1750 m depth by Middle Oligocene time (Sclater et al., 1971). The oldest sediments are of Late Eocene age overlying basement basalts.

The nannoplankton were initially studied by Ellis (1975) and his zonation agrees well with the other deep-sea sites studied when compared with the planktic foraminiferal record. Ujiie (1975) provided an initial biostratigraphic report on planktic foraminifers.

Faunal associations and abundance changes in Site 292 are most similar to those of the Indian Ocean Site 219 and South Atlantic Site 363 during the Late Eocene and Early Oligocene. The Oligocene sequence is also very similar to the eastern equatorial Pacific Site 77B where only very latest Eocene sediments are present. Hence, faunal correlations can be made with the low latitude Indian and Atlantic Oceans.

At Site 292 basement rocks underlie the basal *G. semiinvoluta* Zone, an interval which in Sites 219, 363 and 253 is marked by a
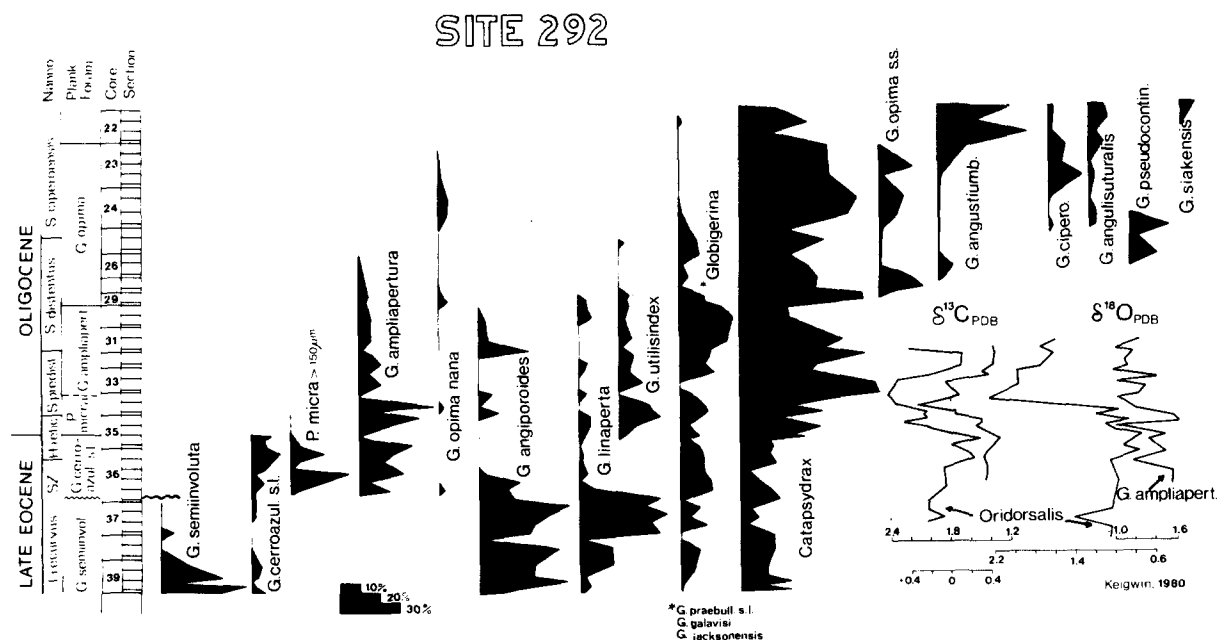

Fig. 9. Abundance distributions of planktic foraminiferal species of Site 292, northwestern Pacific. Wavy line marks hiatus. Nannoplankton zonation after Ellis (1975). Isotope data after Keigwin (1980).
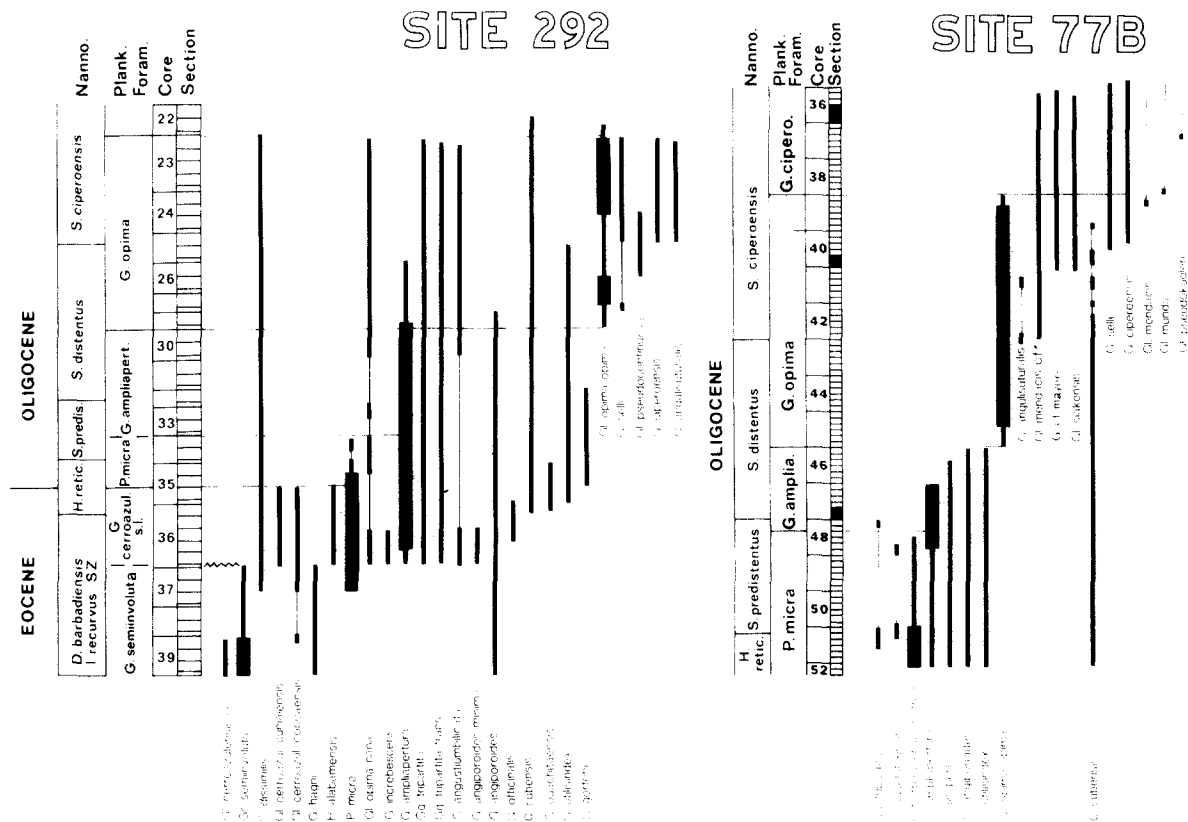
Fig. 10. Ranges of selected planktic foraminiferal species of Sites 292 and 77B, northwestern and eastern equatorial Pacific. Thick lines mark relative abundance of index species; wavy line marks hiatus. Nannoplankton zonation in Site 292 after Ellis (1975), in Site 77B after Bukry (1972).

hiatus. The latest Eocene hiatus which marks the *G. semiinvoluta/G. cerroazulensis* Zone boundary in Sites 219 and 363 is also present in Site 292. As at these sites the upper limit of this hiatus appears to be in the lower half of the *G. cerroazulensis* Zone (P16) as indicated by the appearance of *G. increbescens*, *G. ampliapertura*, *G. cerroazulensis cunialensis* and other species at this interval (Fig. 9). Just as in Site 219, cooler water species *G. angiporoides* (30—40%), *G. linaperta* (20—40%) and *G. semiinvoluta* dominate the *G. semiinvoluta* Zone; however, *P. micra* (>150 µm) is rare (Fig. 10). The *G. cerroazulensis* Zone is characterized by abundant *P. micra* >150 µm, *G. ampliapertura* and common *G. ouachitaensis* which also dominate this interval at Site 219 (Figs. 6, 10).

The Eocene/Oligocene boundary, based on the extinction of *Hantkenenia alabamensis*

and the *G. cerroazulensis* group, falls within the *H. reticulata* nannoplankton Zone as at Sites 219, 363 and 77B (Fig. 7). Contrary to the commonly held view that an unconformity characterizes the E/O boundary, no hiatus has been observed in any of these sites. In order to take a closer look at the geological record at this boundary in a sedimentary sequence that appears to be continuous, an expanded view of Site 292 is illustrated in Fig. 9 along with the oxygen isotope curve of Keigwin (1980). Note that three intervals of increased carbonate dissolution (diagonal lines, Fig. 9) are present at the latest Eocene hiatus, at the E/O boundary, and in the lower part of the *G. ampliapertura* Zone. Each of these intervals corresponds to generally higher $\delta^{18}O$ values, indicating cooling, and higher $\delta^{13}C$ values which, in the modern ocean are often associated with younger water masses

(Miller and Curry, 1982). This suggests that carbonate dissolution was caused by increased production of young water during periods of increased polar refrigeration.

The sudden drop in $\delta^{18}O$ inferred bottom-water temperatures accompanied by a lesser cooling in $\delta^{18}O$ inferred surface water temperatures, starts in the latest Eocene and reaches the maximum low in the Early Oligocene, in the upper half of the *P. micra* Zone. Unfortunately, the exact timing of this maximum cooling can not be determined due to incomplete core recovery. The faunal response to these temperature fluctuations is somewhat biased due to selective dissolution during intervals of increased dissolution. The species most affected by dissolution are *P. micra* and *G. ampliapertura* (Fig. 11; see also Site 219, Fig. 6). *Catapsydrax* is dissolution-resistant and thus enriched during dissolution pulses. Although dissolution is an

obscuring factor, it is obvious that paleo-ecologic conditions were less favorable for *G. cerroazulensis*, *P. micra*, *G. ampliapertura*, *G. linaperta*, and *G. angioporoides* by the earliest Oligocene cooling whereas *Catapsydrax* flourished (Fig. 11).

Bottom temperatures temporarily increased whereas surface temperatures remained stable during the *G. ampliapertura* Zone. An increase in *Globigerina praebulloides*, *G. galavisi*, and slight decline in *Catapsydrax* marks this interval. A drastic faunal change occurs within the *G. opima* Zone. The dominant faunal constituents in this zone are *G. opima opima*, *G. pseudocontinuosa*, *Globigerina angustiumbilicata* and *G. ciperoensis* (Fig. 10) just as at Sites 363, 77B and 253.

Site 77B

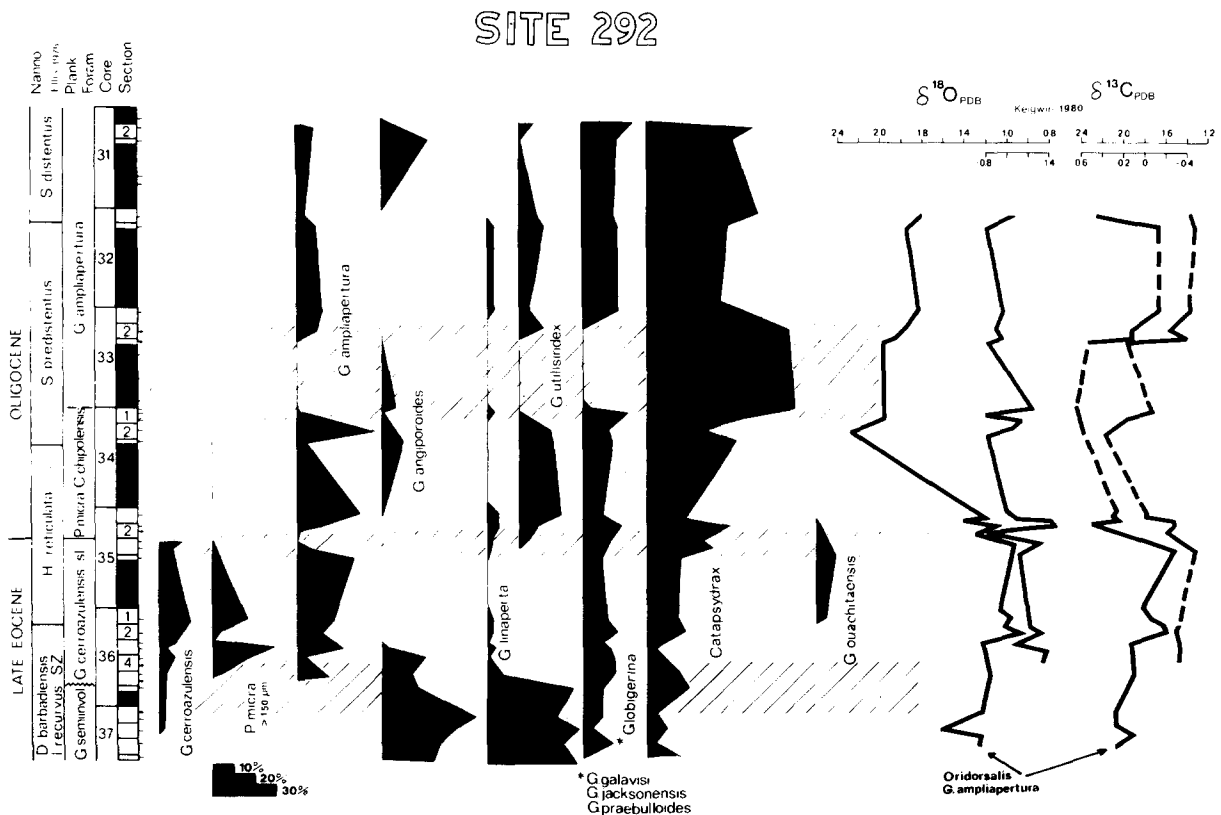The equatorial Pacific Site 77B (0°28.9'N

Fig. 11. Abundance distributions of dominant planktic foraminiferal species and isotope data (Keigwin, 1980) across the Eocene/Oligocene boundary at Site 292. Diagonal lines mark intervals of increased dissolution.
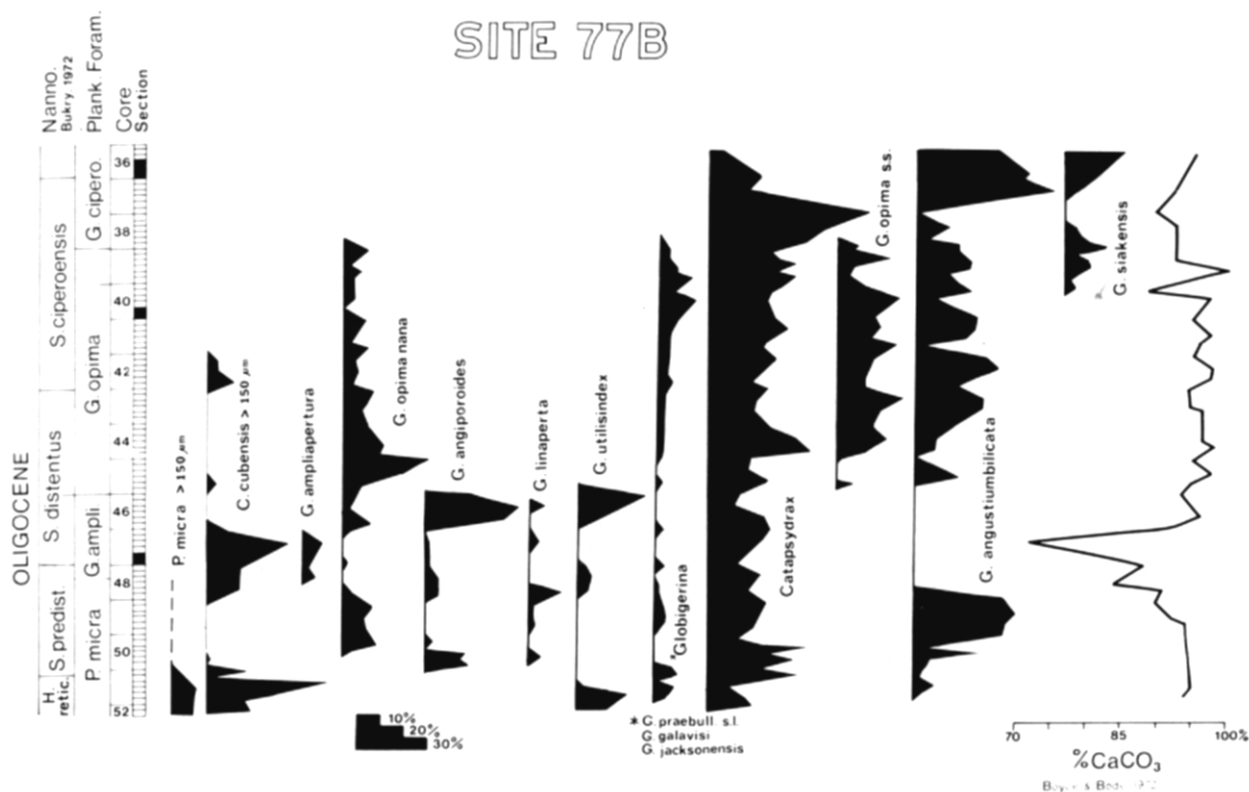
Fig. 12. Abundance distributions of dominant planktic foraminiferal species in Site 77B.

133°13.7'W) was cored at 4290 m water depth. The oldest sediment overlying oceanic basement is of Late Eocene age (Hays et al., 1972). During the Late Eocene deposition occurred at 2200 m depth subsiding to 3100 m by Middle Oligocene time (Sclater et al., 1971). Throughout its depositional history, Site 77B remained within a few degrees of the Equator and above or near the carbonate compensation depth.

Nannoplankton stratigraphy was initially reported by Bukry (1972) and Gartner (1972). Jenkins and Orr (1972) studied the planktic foraminifers and erected a new zonation claiming that dissolution made Bolli and Blow Zones inapplicable. Although dissolution effects at Site 77B were observed to be stronger than at Site 292, it is possible to apply Bolli's zonation.

Planktic foraminifers are absent whereas nannoplankton are common in the chalk and clay sediments deposited during the latest Eocene *Discoaster barbadiensis* nannoplankton Zone (Bukry, 1972). The earliest planktic foraminiferal assemblages are found in Core 52-2 and are characterized by abundant *P. micra* >150 μm and *Chiloguembelina cubensis* >150 μm which decline in abundance near the top of Core 51 (Figs. 9, 12). No Upper Eocene index species are present. Faunal correlations with other sites indicate that this assemblage occurs at, or near, the E/O boundary. For instance, in Sites 363, 319 and 292 the last common *P. micra* >150 μm occurs at, or slightly above, the E/O boundary. Another secondary marker for the earliest Oligocene *P. micra* Zone, *Globigerina utilisindex* which first appears in abundance near the E/O boundary in Sites 363 and 292, is also present in Site 77B. The presence of common *P. micra* (>150 μm) and *G. utilisindex* in Core 52-2, therefore suggests that this interval is at or near the E/O boundary. This interval also correlates to the

base of the *H. reticulata* nannoplankton Zone, and the same relative position of the two zones is observed in other sites. The major cooling observed in the benthic oxygen isotope curve occurs in Core 50 (Keigwin, unpublished data), coincident with the disappearance of *P. micra* (>150 µm) (also *C. cubensis* >150 µm, Fig. 12) as also observed in Sites 292 and 363.

Faunal assemblages and abundance fluctuations at Site 77B are similar to other low latitude sites with the exception of *C. cubensis* (>150 µm) and *Globigerina angustiumbilicata* which are more abundant in Site 77B. These two species behave antithetically, when *C. cubensis* is abundant, *G. angustiumbilicata* declines (Fig. 12). As at other low latitude sites, the *G. ampliapertura* Zone is characterized by common *G. ampliapertura* and the last common *G. angiporoides*, *G.*

linaperta and *G. utilisindex*. In addition, *C. cubensis* >150 µm is abundant at Site 77B. A carbonate low is present between the upper *P. micra* Zone and the middle of *G. ampliapertura* Zone (Boyce and Bode, 1972) (Fig. 12). As at Sites 292 and 363 the *Globorotalia opima* Zone contains abundant *G. opima opima*, *G. angustiumbilicata*, and *Catapsydrax*.

Site 277

Site 277 was cored in the South Pacific on the southern Campbell Plateau (52° 13.43'S 166°11.48'E) in 1214 m of water. The Middle Eocene to Middle Oligocene nannoplankton ooze and chalk were deposited at 450—650 m depth (Sclater et al., 1971) and about the present paleolatitude of this site (Lowrie and Isrofil, 1975).
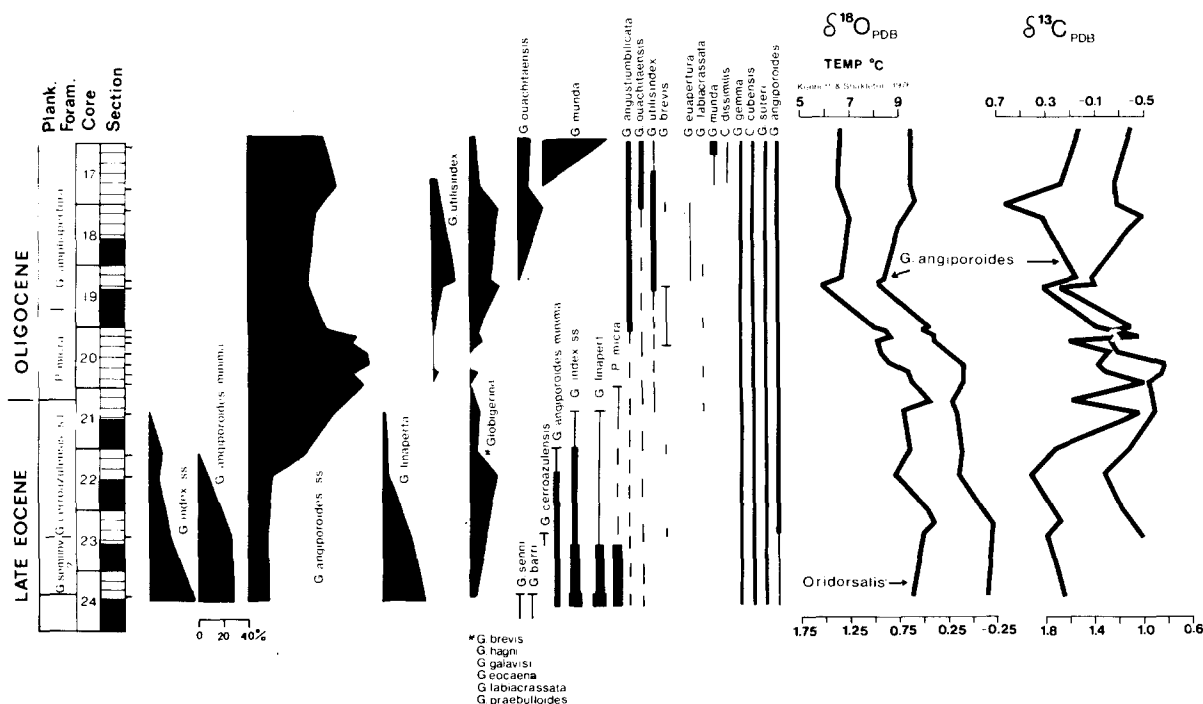

Fig. 13. Abundance distributions of dominant planktic foraminiferal species and ranges of selected species of Site 277, southwestern Pacific. Thick lines in range chart mark relative abundance of species. Tick marks on core-section indicate sample locations.

The high latitude faunal and floral assemblages of Site 277 have necessitated the use of different zonal schemes for both nannoplanktons (Edwards and Perch-Nielsen, 1975) and planktic foraminifers (Jenkins, 1975). A high latitude planktic foraminiferal zonation for the southern oceans was developed by Jenkins based on New Zealand faunal sequences (Jenkins, 1966). A common ground between low and high latitude zonations must be found, however, if we are to correlate across latitudes. This may be achieved based on faunal abundance fluctuations, which in effect may be viewed as the expression of world wide paleoclimatic and paleoceanographic changes affecting faunas across latitudes. The underlying assumption here is that any given paleoclimatic/paleoceanographic change would favor certain species over others which could be detected in the quantitative record. Identification of such faunal changes has been attempted at Site 277 based on comparison of faunal changes as well as the oxygen isotope record observed in Sites 363, 292, 77B and 253.

As noted earlier, the Late Eocene *G. semiinvoluta* Zone is characterized by abundant *P. micra* >150 μm, the last abundance peaks of *Globigerinatheka*, and common *Globigerina linaperta* and *G. angiporoides*. In Site 277 these abundance changes occur in Core 23, where thus the top of *G. semiinvoluta* Zone has been tentatively placed (Fig. 13). This interval corresponds to the *G. linaperta* Zone of Jenkins (1975) and *R. bisecta* nannoplankton Zone of Edwards and Perch-Nielsen (1975). In the interval assigned to the *G. semiinvoluta* Zone, it is notable that the thin walled *G. angiporoides minima* is common while the larger thick walled *G. angiporoides* dominates later. In low latitude sequences only thin walled *G. angiporoides* are present, but they are not typical of *G. angiporoides minima* and hence are not differentiated.

Another possible interpretation places the upper limit of the *G. semiinvoluta* Zone at the E/O boundary suggesting the presence of a hiatus at this interval similar to Site 253. This interpretation is based on the range of *Globigerinatheka index* which is not observed

in the low latitude sites studied above the *G. semiinvoluta* Zone. The absence of faunal abundance indices of the *G. cerroazulensis* Zone (*P. micra* > 150 μm, *G. ampliapertura*, and *G. ouachitaensis*) in Site 277 also points towards the presence of a hiatus. It will be necessary to examine further high latitude deep-sea sequences to determine the completeness of this interval in Site 277. In this paper it is assumed that Site 277 does not contain a hiatus, hence the E/O boundary marks the upper limit of the *G. cerroazulensis* Zone.

The E/O boundary has been tentatively placed between Cores 20cc and 21-3 based on the first appearance of *Globigerina utilisindex* and initial decline in water temperatures as indicated by oxygen isotope data (Fig. 13). This conclusion has been reached by comparison with Sites 292, 253 and 77B where *G. utilisindex* first appears in the Late Eocene and is common in the latest Eocene to earliest Oligocene coincident with the decline in water temperatures as indicated by oxygen isotope data (Fig. 13). At Site 277 this boundary is in close agreement with the E/O boundary as determined by nannofossil biostratigraphy (Edwards and Perch-Nielsen, 1975).

The *P. micra* and *G. ampliapertura* Zones can be recognized by the range of *G. utilisindex* which disappears near the *G. ampliapertura/G. opima* Zone boundary (see also Sites 292, 253, 77B). The presence of common *G. ouachitaensis* is a secondary marker for the *G. ampliapertura* Zone (Fig. 13). The boundary between *P. micra/G. ampliapertura* has been placed between Cores 19 and 20 based on the maximum cooling observed in the oxygen isotope record.

Biostratigraphic synthesis

Biostratigraphic and faunal analyses of Middle Eocene to Oligocene DSDP Sites 363, 219, 292 and 77B demonstrate that Bolli's (1957a, b, 1966, 1972) zonation can be applied to low latitude sequences in the Atlantic, Pacific and Indian oceans. Moreover, species abundance distributions follow

the same general patterns in all three oceans and their abundance changes further aid in biostratigraphic correlation. The combined use of planktic foraminiferal zones, datum events, and species abundance changes provides a high resolution chronostratigraphy for Eocene to Oligocene low latitude sequences (Fig. 2).

Planktic foraminiferal zones in low latitudes can be characterized by the abundance changes of the most dominant species and hence biostratigraphic zones can be inferred even in the absence of zonal index species. For instance, the faunal assemblage of the *Globigerinatheka subconglobata* Zone (P11) is dominated by *Globorotalia broedermanni* which rapidly declines into the *G. lehneri* Zone (P12). A greater diversity of species occurs in the *G. lehneri* Zone, as for example *G. lehneri*, *G. spinulosa*, *G. bullbrooki* and increasingly common *T. rohri*. This latter species reaches its apex in the *Truncorotaloides rohri* Zone (P14). The dominant species of the *Globigerina semiinvoluta* Zone (P15) are *G. semiinvoluta* and the first abundance peak of *P. micra* >150 μm. A second and last abundance peak of *P. micra* >150 μm and common *G. ouachitaensis* (in the Indian Ocean) marks the *G. cerroazulensis* Zone (P16-P17) along with increasing abundances of *G. linaperta* and *G. angiporoides*. The Early Oligocene *P. micra* Zone (P18-P19) and *G. ampliapertura* Zone (P20) show increasing abundances of *Catapsydrax*, *Globigerina galavisi* and *G. praebulloides* along with common *G. ampliapertura*. A characteristic faunal change occurs in the *G. opima* Zone with the appearance and dominance of *G. opima opima*, *G. pseudocontinuosa* and *Globigerina angustiumbilicata*. Recognition of these faunal assemblage changes provides an additional tool for biostratigraphic correlation.

Quantitative foraminiferal analysis of the high latitude South Pacific Site 277 suggests that a stratigraphic correlation between high and low latitude deep-sea sequences may be possible based on population changes of planktic foraminifers. At this time, however, only tentative correlations can be made and

further study of mid and high latitude sections are necessary to develop this correlation method.

Two hiatuses occurring at Middle/Late Eocene boundary and in the latest Eocene *G. semiinvoluta/G. cerroazulensis* Zone boundary appear to have a global distribution. Detailed biostratigraphic study of further deep-sea sections may reveal a third hiatus in the Middle Eocene *Orbulinoides beckmanni* Zone as suggested in Site 219. There is no evidence of a hiatus at the E/O boundary in low and middle latitude sequences. A hiatus may also be present in the Late Oligocene *Globorotalia opima* Zone, as suggested by the abrupt assemblage changes in Site 253 at this interval. No hiatus appears to be present at this interval in low latitude sequences.

In the highly productive equatorial regions the Middle/Late Eocene boundary hiatus and the latest Eocene hiatus are short, whereas in high latitude sections longer hiatuses may be expected perhaps due to the greater intensity and erosiveness of bottom currents as suggested at the Indian Ocean Site 253. The presence of reworked older faunas at these hiatus intervals in many deep-sea sites suggests that the hiatuses are caused by erosion, rather than nondeposition due to increased dissolution.

Paleoclimatic interpretation

Population studies of planktic foraminifers of the most continuous sequences in the Atlantic, Pacific and Indian oceans provide evidence of the global impact of the paleoclimatic, and paleoceanographic changes occurring during the Middle Eocene to Oligocene time. The high resolution biochronology and paleoclimatic information based on temperature sensitive species abundance data permit more accurate placing of these events in the geologic time frame.

A global paleoclimatic and paleoceanographic signal can be recognized by five major planktic foraminiferal assemblage changes indicating major cooling episodes during Middle Eocene to Oligocene time: Middle Eocene (P13) 46—45 Ma, Middle/Late Eocene

boundary (P14/P15) 44—43 Ma, Late Eocene (P15/P16) 41—40 Ma, earliest Oligocene (P18) 37—36 Ma, and Late Oligocene (P20/ P21) 31—29 Ma (Fig. 2). Cooling episodes at these times, except for P13 and P15/P16, have also been observed in planktic faunas and floras by Haq et al. (1977). Oxygen isotope studies also indicate cooling episodes at these times with the exception of 46—45 Ma (P13) and 31—29 Ma (P20/P21) where the data is inconclusive (Shackleton and Kennett, 1975; Keigwin, 1980; Miller and Curry, 1982). Three of these cooling episodes, at the Middle/Late Eocene boundary (P14/ P15), the Late Eocene (P16), and Late Oligocene (P21), correlate to eustatic sea level low stands (Vail et al., 1977).

Middle to Late Eocene

Major faunal assemblage changes may be viewed as faunal migrations poleward and equatorward in response to climatic cooling and warming as also interpreted from nanno-plankton assemblages (Haq and Lohmann, 1976). The present faunal analyses indicate that generally warm, but fluctuating, climatic conditions prevailed during the Middle Eocene (P11—P12) with spinose species (*Globorotalia broedermanni* and *G. bullbrooki*) dominating in the tropics. A faunal change occurs in Zone P13 (44—43 Ma) resulting in the decline of *G. broedermanni* and *G. bullbrooki* and dominance of *Truncorotaloides rohri*. A hiatus is present at this interval at Site 219, and at Site 363 Zone P13 is very short (Core 10cc), suggesting non-deposition due to increased carbonate dissolution, or erosion due to increased bottom-water circulation. No oxygen isotope data is available for this interval.

A major faunal change, indicating cooling, occurs at the Middle/Late Eocene boundary (P14/P15), resulting in the extinction of the spinose tropical fauna and equatorward migration of mid-latitude faunas with cooler water masses. These mid-latitude faunas (consisting of *Globorotalia carcosellensis*, *Globigerina medizzai*, and *Globigerinatheka*) survived in equatorial regions through the

early Late Eocene, only to become extinct during another cool interval in the Late Eocene (P15/P16) and replaced by the equatorward migration of a still cooler mid- to high-latitude fauna dominated by *Globigerina angiporoides*, *G. linaperta* and *G. ampliapertura*. The two major faunal changes, at the Middle/Late Eocene (P14/P15) and Late Eocene (P15/P16), are associated with isotopic cooling (Keigwin, 1980; Miller and Curry, 1982), and widespread deep-sea hiatuses in high- and low-latitude sections studied. This suggests major increases in the production of bottom-water and intensified current circulation at these times.

Eocene/Oligocene Boundary

Oxygen isotope data indicate that the major cooling in bottom-water temperatures, and high latitude surface water temperatures, occurred at the E/0 boundary and into the Early Oligocene (P18) (Shackleton and Kennett, 1975; Savin et al., 1975; Keigwin, 1980; Miller and Curry, 1982). Few planktic foraminiferal species extinctions, however, occur at this time in either high latitude or tropical sequences. The faunal change is apparent primarily in the increased abundance of cool water species (*Globigerina linaperta*, *G. utilisindex*, *G. angustiumbilicata*, *G. galavisi* and *Catapsydrax*). Corliss (1981) observed a similar lack of major faunal change in benthic foraminiferal faunas at this inferred bottom-water event. He thus suggests that the $\delta^{18}O$ enrichment near the E/0 boundary is, in part, due to rapid ice accumulation as proposed by Mathews and Poore (1980). With this assumption, the inferred temperature change can be calculated to be about 2°—3°C (Corliss, 1981; Miller and Curry, 1982). This temperature change is well within the environmental tolerance of benthic species and, hence, could explain the absence of a major faunal change at this time. Planktic foraminiferal faunas also suggest either a greater environmental tolerance, or a less significant surface water cooling, than during the earlier two cooling episodes (P14/P15, and P15/P16).

Late Oligocene

The fifth major faunal change, and cooling episode, occurs in the Late Oligocene 31—29 Ma (P20/P21), and is associated with a major drop in the eustatic sea level (Vail and Hardenbol, 1979). This faunal change is marked by the extinction of many cool water species which dominated during the Early Oligocene (*G. linaperta*, *G. angiporoides*, *G. utilisindex*, *G. ampliapertura*). Other common cool water Early Oligocene species (*G. angustiumbilicata*, *Catapsydrax*), however, continue to dominate along with *Globorotalia opima opima* and *G. pseudocontinuosa*. A cooling phase at this time has also been observed in nannoplankton assemblages (Haq et al., 1977). Published oxygen isotope data, however, do not show a significant temperature change at this interval (Shackleton and Kennett, 1975; Savin et al., 1975; Miller and Curry, 1982).

The fact that a similar series of major cooling episodes is implied by the oxygen isotope record, the faunal and floral data, and the sea level curve indicates that these factors are likely to respond to the same environmental input. The Middle Eocene to Oligocene cooling episodes can be related to continental drift, the separation of Australia from Antarctica, and the development of the Circum-Antarctic Current. In this scenario, the Middle Eocene cooling (P13, 44—43 Ma) is probably due to partial thermal isolation of Antarctica resulting from northward drift of Australia. The major cool event at the Middle/Late Eocene boundary, affecting a major faunal change, isotopic cooling, widespread hiatus formation, and lowering of the eustatic sea level, most likely records the initial development of the prototype of the Circum-Antarctic current. This development presumably led to further Antarctic cooling culminating in a major cool event associated with a faunal change and lower eustatic sea level in the Late Eocene (P15/P16). Hiatus formation is widespread at this time in both low and high latitudes and presumably resulted from intensified bottom current circulation. The increasing abundances of cool water species and absence of major species extinctions and first appearances, the change in sea level, and absence of a hiatus at the E/O boundary, suggest that the major enrichment in $\delta^{18}O$ and development of the psychrosphere resulted from continued cooling, and/or the rapid build-up of a major Antarctic ice sheet, as suggested by Matthews and Poore (1980), Corliss (1981), and Miller and Curry (1982).

The faunal cooling observed in the Late Oligocene is associated with a major sea level drop, but current data does not show a significant isotopic cooling. Further isotopic studies are needed to interpret this faunal change.

Conclusions

Planktic foraminiferal assemblages have been analyzed quantitatively in six DSDP Sites (363, 292, 77B, 277, 219, 253) in order to determine the nature of the faunal turnover during Middle Eocene to Oligocene time. Biostratigraphic ranges of taxa and abundance distributions of dominant species are presented and illustrate: (1) striking similarities in faunal assemblages of low latitude regions in the Atlantic, Pacific and Indian oceans; (2) application of Bolli's zonal scheme is possible in the mid- and low-latitude regions of all three oceans, thus permitting easy correlation between oceans; (3) 55 datum levels are found to be isochronous between Zones P11 and P20 and provide a high resolution biochronology; (4) planktic foraminiferal zones are characterized by specific species abundances reflecting environmental conditions.

Population studies reveal 5 major faunal changes indicating cooling episodes: at the Middle Eocene (P13) 44—43 Ma, the Middle/Late Eocene boundary (P14/P15) 41—40 Ma, the Late Eocene (P15/P16) 39—38 Ma, the Eocene/Oligocene boundary (P18) 37—36 Ma and the Late Oligocene (P20/P21) 31—29 Ma. Three of these faunal changes are associated with isotopic cooling (P14/P15,

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P15/P16, P18). These faunal changes indicate the replacement of a warm Middle Eocene fauna by successively cooler faunal assemblages indicating a gradual cooling occurred between Middle Eocene and Oligocene time. The faunal change at the E/O boundary (P18), associated with the major inferred bottom-water cooling, is evident in increased abundances of cool water species, rather than major extinctions and first appearances as during the other cooling episodes.

Four hiatus intervals have been recognized in Zones P13, P14/P15, P15/P16, and P20/P21, corresponding to intervals of major faunal changes. No hiatus has been observed at the E/O boundary. Hiatuses at the Middle/Late Eocene boundary and in the Late Eocene (P15/P16) are widespread in low and high latitude sequences. These hiatuses suggest that vigorous bottom-water circulation began developing in the Middle Eocene, well before the development of the psychrosphere at the E/O boundary.

References

Beckmann, J.P., Bolli, H.M., Perch-Nielsen, K., Proto Decima, F., Saunders, J.B. and Toumarkine, M., 1981. Major calcareous nannofossil and foraminiferal events between the middle Eocene and early Miocene, Paleogeogr., Paleoclimatol., Paleoecol., 36(3/4): 155—190.

Berger, W.H., 1972. Deep-sea carbonates: dissolution facies and age depth constancy. Nature, 236: 392—395.

Berger, W.H. and Winterer, E.L., 1974. Plate stratigraphy and the fluctuating carbonate line. Spec. Pub. Int. Assoc. Sedimentol., 1: 11—48.

Berggren, W.A., 1972. A Cenozoic time scale — some implications for regional geology and paleobiogeography. Lehtaia, 5(2): 195—215.

Berggren, W.A., 1978. Recent advances in Cenozoic planktonic foraminiferal biostratigraphy, biochronology, and biogeography, Atlantic Ocean. Micropaleontology, 24(4): 337—370.

Blow, W.H., 1969. Late Middle Eocene to Recent planktonic foraminiferal biostratigraphy. In: Proc. Plankt. Conf. Geneva, 1st, 1967, pp. 199—422.

Bolli, H.M., 1957a. Planktonic foraminifera from the Oligocene—Miocene Cipero and Lengua Formations of Trinidad, B.W.I. U.S. Natl. Mus. Bull., 215: 155—172.

Bolli, H.M., 1957b. Planktonic foraminifera from the Eocene Navet and San Fernando formations of Trinidad, B.W.I. U.S. Natl. Mus. Bull., 215: 155—172.

Bolli, H.M., 1966. Zonation of Cretaceous to Pliocene marine sediments based on planktonic foraminifera. Bol. Inf. Asoc. Venez. Geol. Min. Petrol., 9: 3—31.

Bolli, H.M., 1972. The genus Globigerinatheka Broenimann. J. Foraminiferal Res., 2(3): 109—136.

Boltovskoy, E., 1974. Neogene planktonic foraminifera of the Indian Ocean. In: Initial Reports of the Deep Sea Drilling Project, 26: 675—742.

Boudreaux, J.E., 1974. Calcareous nannoplankton ranges, DSDP Leg 23. In: Initial Reports of the Deep Sea Drilling Project, 23: 1073—1090.

Boyce, R.E. and Bode, G.W., 1972. Carbon and carbonate analyses — Leg 9, DSDP. In: Initial Reports of the Deep Sea Drilling Project, 9: 797—816.

Brunet, M., 1979. Les grands Mammifères chefs de file de l'imigration Oligocène et le problème de la limite Eocène—Oligocène en Europe. Foundation Singer-Plignac, Paris, 223 pp.

Bukry, D., 1972. Coccolith stratigraphy, Leg 9 DSDP. In: Initial Reports of the Deep Sea Drilling Project, 9: 817—832.

Bukry, D., 1973. Low-latitude coccolith biostratigraphic zonation. In: Initial Reports of the Deep Sea Drilling Project, 15: 685—703.

Bukry, D., 1974. Coccolith stratigraphy, Arabian and Red Seas, DSDP Leg 23. In: Initial Reports of the Deep Sea Drilling Project, 23: 1091—1094.

Bukry, D., 1978. Cenozoic silicoflagellate and coccolith stratigraphy, southeastern Atlantic Ocean, DSDP Leg 40. In: Initial Reports of the Deep Sea Drilling Project, 40: 635—650.

Cavalier, C., 1979. La limite Eocène—Oligocène en Europe occidentale. Memo Sci. Geol., 54: 280 pp.

Corliss, B.H., 1981. Deep-Sea benthonic foraminiferal faunal turnover near the Eocene/Oligocene boundary. Mar. Micropaleontol., 6: 367—384.

Donn, W.L. and Shaw, D.M., 1977. Model of climatic evolution based on continental drift and polar wandering. Geol. Soc. Am. Bull., 88: 390—396.

Douglas, R.G. and Savin, S.M., 1975. Oxygen and carbone isotope analyses of Tertiary and Cretaceous microfossils from Shatsky Rise and other sites in the North Pacific Ocean. In: Initial Reports of the Deep Sea Drilling Project, 32: 509—520.

Edwards, A.R. and Perch-Nielsen, K., 1975. Calcareous nannofossils from the southern southeast Pacific, DSDP Leg 29. In: Initial Reports of the Deep Sea Drilling Project, 29: 469—487.

Ellis, C.H., 1975. Calcareous nannofossil biostratigraphy, Leg 31, DSDP. In: Initial Reports of the Deep Sea Drilling Project, 31: 655—692.

Fleisher, R.L., 1974. Cenozoic planktonic foraminifera and biostratigraphy, Arabian Sea DSDP Project, Leg 23. In: Initial Reports of the Deep Sea Drilling Project, 23: 1001—1072.

Gartner, S. Jr., 1972. Coccolith age determinations, Leg 9, DSDP. In: Initial Reports of the Deep Sea Drilling Project, 9: 833—840.

Hamilton, W., 1968. Cenozoic climatic change and its cause. Meteorol. Monographs, 8: 128—133.

Haq, B.L. and Lohmann, G.P., 1976. Early Cenozoic calcareous nannoplankton biogeography of the Atlantic Ocean. Mar. Microapleontol., 1: 119—194.

Haq, B.U., Premoli-Silva, I. and Lohmann, G.P., 1977. Calcareous plankton paleobiogeographic evidence for major climatic fluctuations in the early Cenozoic Atlantic Ocean. J. Geophys. Res., 82(27): 3861—3876.

Hardenbol, J. and Berggren, W.A., 1978. A new Paleogene numerical time scale. AAPG Stud. Geol., 6: 213—234.

Hays, J.D., Cook, H.E., Jenkins, D.G., Cook, F.M., Fuller, J.T., Goll, R.M., Milow, D.E. and Orr, W.N., 1972. Site 77. In: Initial Reports of the Deep Sea Drilling Project, 9: 43—208.

Jenkins, D.G., 1966. Planktonic foraminiferal zones and new taxa from the Danian to lower Miocene of New Zealand. N.Z.J. Geol. Geophys., 8: 1088—1122.

Jenkins, D.G., 1975. Cenozoic planktonic foraminiferal biostratigraph of the southwestern Pacific and Tasman Sea — DSDP Leg 29. In: Initial Reports of the Deep Sea Drilling Project, 29: 449—467.

Jenkins, D.G. and Orr, W.N., 1972. Planktonic foraminiferal biostratigraph of the eastern equatorial Pacific, Leg 9. In: Initial Reports of the Deep Sea Drilling Project, 9: 1095—1196.

Keigwin, L.D. Jr., 1980. Paleoceanographic change in the Pacific at the Eocene—Oligocene boundary. Nature, 287(5784): 722—725.

Keller, G., 1980. Middle to Late Miocene planktonic foraminiferal datum levels and paleoceanography of the north and southeastern Pacific Ocean. Mar. Micropaleontol., 5: 249—281.

Kennett, J.P., 1977. Cenozoic evolution of Antarctic glaciation, the Circum-Antarctic Ocean, and their impact on global paleoceanography. J. Geophys. Res., 82(27): 3843—3859.

Kennett, J.P., 1978. The development of planktonic biogeography in the southern ocean during the Cenozoic. Mar. Micropaleontol., 3: 301—345.

Kennett, J.P., Burns, R.E., Andrews, J.E., Churkin, M., Davies, T.A., Dumitrica, P., Edwards, A.R., Galehouse, J.S., Packham, G.H. and Van der-Lingen, J.G., 1972. Australian—Antarctic continental drift, paleocirculation changes and Oligocene deep-sea erosion. Nature Phys. Sci., 239: 51.

Kennett, J.P., Houtz, R.E., Andrews, P.B., Edwards, A.R., Gostin, V.A., Hajos, M., Hampton, M.A., Jenkins, D.G., Margolis, S.V., Ovenshine, A.T. and Perch-Nielsen, K., 1975. Cenozoic paleoceanography in the southwest Pacific Ocean, Antarctic Glaciation and the development of the circum-Antarctic Current. In: Initial Reports of the Deep Sea Drilling Project, 29: 1155—1170.

Kidd, R.B. and Davies, T.A., 1978. Indian Ocean sediment distribution since the Late Jurassic. Mar. Geol., 26: 49—70.

Loeblich, A.R. Jr. and Tappan, H., 1957. Planktonic Foraminifera of Paleocene and early Eocene age from Gulf and Atlantic Coastal Plains. B.W.I. U.S. Natl. Mus. Bull., 215: 173—198.

Lowrie, W. and Isrofil, N.M., 1975. Paleomagnetism of basalt samples from Leg 29. In: Initial Reports of the Deep Sea Drilling Project, 29: 1109—1116.

Martini, E., 1971. Standard Tertiary and Quaternary calcareous nannoplankton zonation. Proc. Plankt. Conf., 2nd, pp. 739—785.

Matthews, R.K. and Poore, R.Z., 1980. Tertiary ^{18}O record and glacio-eustatic sea-level fluctuations. Geology, 8: 501—504.

McGowran, B., 1978. Stratigraphic record of early Tertiary oceanic and continental events in the Indian Ocean region. Mar. Geol., 26: 1—39.

McKenzie, D. and Sclater, J.G., 1971. The evolution of the Indian Ocean since the late Cretaceous. R. Astron. Soc. Geophys. J., 25: 437—528.

Melguen, M., 1978. Facies evolution, carbonate dissolution cycles in sediments from the eastern South Atlantic (DSDP Leg 40) since the Early Cretaceous. In: Initial Reports of the Deep Sea Drilling Project, 40: 981—1024.

Miller, K.G. and Curry, W.B., 1982. Eocene to Oligocene benthic foraminiferal isotopic record in the Bay of Biscay. Nature, 296(5855): 347—352.

Moore, T.C. Jr., Van Andel, T.H., Sancetta, C. and Pisias, N., 1978. Cenozoic hiatuses in marine sediments. Micropaleontology, 24: 113—138.

Pierce, J.W., Denham, C.R. and Luyendyk, B.P., 1974. Paleomagnetic results of basalt samples from DSDP Leg 26, South Indian Ocean. In: Initial Reports of the Deep Sea Drilling Project, 26: 517—527.

Proto Decima, F., Medizza, F. and Todesco, L., 1978. Southeastern Atlantic Leg 40, Calcareous nannofossils. In: Initial Reports of the Deep Sea Drilling Project, 40: 571—634.

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Ryan, W.B.F., Bolli, H.M., Foss, G.N., Natland, H.J., Hottman, W.E. and Foresman, J.B., 1978. Walvis Ridge — Sites 362 and 363. In: Initial Reports of the Deep Sea Drilling Project, 40: 183—356.

Sancetta, C., 1979. Paleogene Pacific microfossils and paleoceanography. Mar. Micropaleontol., 4: 363—398.

Savin, S.M., Douglas, R.G. and Stehli, F.G., 1975. Tertiary marine paleotemperatures. Bull. Geol. Soc. Am., 86: 1499—1510.

Sclater, J.G., Anderson, R. and Bell, G., 1971. The evolution of ridges and the evolution of the central eastern Pacific. J. Geophys. Res., 76: 7888—7915.

Sclater, J.G., Abbott, D. and Thiede, J., 1977. Paleobathymetry and sediments of the Indian Ocean. In: J.R. Heirtzler et al. (Editors), Indian Ocean Geology and Biostratigraph. Am. Geophys. Union, Washington, D.C., pp. 25—29.

Shackleton, N.J. and Kennett, J.P., 1975. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: oxygen and carbon isotope analyses in DSDP Sites 277, 279, and 281. In: Initial Reports of the Deep Sea Drilling Project, 29: 743—755.

Stainforth, R.M., Lamb, J.L., Luterbacher, H., Beard, J.H. and Jeffords, R.M., 1975. Cenozoic Planktonic Foraminiferal Zonation and Characteristics of Index Forms. The University of Kansas Paleontological Contribution, the University of Kansas, Lawrence, Kansas, 425 pp.

Stehlin, H.G., 1909. Remarques sur les faunules de Mammifères des couches éocènes et oligocène du Bassin de Paris. Bull. Soc. Geol. Fr., (4), 18: 488—520.

Thierstein, H.R., 1974. Calcareous nannoplankton — Leg 26, DSDP. In: Initial Reports of the Deep Sea Drilling Project, 26: 619—668.

Toumarkine, M., 1978. Planktonic foraminiferal biostratigraph of the Paleogene of Sites 360 to 364 and the Neogene of Sites 362A, 363, and 364, Leg 40. In: Initial Reports of the Deep Sea Drilling Project, 40: 679—722.

Toumarkine, M. and Bolli, H.M., 1970. Evolution de Globorotalia cerroazulensis (Cole) dans l'Eocène moyen et supérieur de Possagno (Italie). Rev. Micropaleontol., 3: 131—145.

Toumarkine, M. and Bolli, H.M., 1975. Foraminifères planctoniques de l'Eocène moyen et supérieur de la Coupe de Possagno. Schweiz. Palontol. Abh., 97: 69—183.

Ujiie, H., 1975. Planktonic foraminiferal biostratigraphy in the western Phillipine Sea, Leg 31, DSDP. In: Initial Reports of the Deep Sea Drilling Project, 31: 677—692.

Vail, P.R. and Hardenbol, J., 1979. Sea level changes during the Tertiary. Oceanus, 22(3): 71—80.

Vail, P.R., Mitchum, R.M. Jr. and Thompson, S., 1977. Global cycles of relative changes of sea level. In: C.E. Payton (Editor), Seismic Stratigraphy, Applications to Hydrocarbon Exploration. Am. Assoc. Petrol. Geol. Tulsa, Okla., pp. 83—98.

Van Andel, T.H., Heath, G.R. and Moore, T.C. Jr., 1975. Cenozoic history and paleoceanography of the central equatorial Pacific Ocean. Geol. Soc. Am. Mem., 143, 134 pp.

Van Couvering, J.A., Aubry, M.P., Berggren, W.A., Bujak, J.P., Naeser, C.W. and Wieser, T., 1981. The Terminal Eocene Event and the Polish connection. Paleogeogr. Paleoclimatol., Paleoecol., 36(3/4): 321—362.

Whitmarsh, R.B., and shipboard scientific party, 1974. Site 219. In: Initial Reports of the Deep Sea Drilling Project, 23: 35—116.

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