

Middle–late Miocene benthic foraminifera in a western equatorial Indian Ocean depth transect: Paleoceanographic implications

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Received 5 May 2006; received in revised form 6 November 2006; accepted 10 November 2006

Abstract

Middle through upper Miocene (17–5 Ma) benthic foraminiferal faunas (>63 µm) from a depth transect in the western equatorial Indian Ocean (DSDP Site 237, ODP Sites 707, 709 and 710; water depth 1500–3800 m) underwent significant faunal changes at ~14–13 Ma and 11–8 Ma. These faunal changes are not easily interpreted, because of the complex factors controlling benthic foraminiferal distribution and abundance. At ~14–13 Ma, the relative abundance of ‘high-productivity’ taxa (*Bolivina*, *Bulimina*, *Melonis* and *Uvigerina* species), although highly variable, generally increased at the two shallower sites (237 and 707), as did the percentage of *Epistominella exigua* (indicative of seasonal productivity) and *Nuttallides umbonifera* (indicative of corrosive bottom waters) at the two deeper sites (709 and 710); the latter species had a peak in abundance at the deepest site (710) between 11 and 9 Ma. Benthic foraminiferal accumulation rates (BFARs) increased strongly at the shallower sites starting at 11 Ma, peaking between 9 and 8 Ma, and increased weakly at the deeper two sites starting at 11 Ma. Elongate, cylindrical taxa decreased in abundance between 12 and 11 Ma at all sites. The abundance of planktic foraminiferal fragments was, as expected, overall higher at the deeper sites, with some high values after 12–13 Ma at the shallower sites. The faunal changes suggest that overall the food supply to the sea floor increased, but also became more pulsed or seasonal, peaking at 9–8 Ma, while deep and intermediate waters may have become more ventilated. The first phase of faunal change (14–13 Ma) was coeval with the global increase in benthic foraminiferal oxygen isotope values, worldwide turnover in benthic foraminiferal faunas, and possibly increased production of Antarctic Bottom Water (AABW) and Northern Component Water (NCW). The overall faunal turnover may reflect the composite faunal response to increased food input resulting from heightened productivity and increased bottom water ventilation associated with a combination of monsoonal intensification, global cooling and changes in ocean circulation, but there is no unequivocal evidence for the presence of Tethyan Outflow Water (TOW) over the studied time and depth range, in contrast to earlier publications.

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Keywords: Miocene; Benthic foraminifera; Indian Ocean; Paleoceanography; Monsoon; ODP

1. Introduction

The Miocene was a time of major changes in the ocean–atmosphere system. By early Miocene times

(~22 Ma) the ocean basins had assumed their modern configurations and the separate Antarctic continent was established (e.g., Ramstein et al., 1997). The East Antarctic Ice Sheet (EAIS), and hence steep thermal gradients between high and low latitudes, probably became established during the earliest Oligocene (~33.5 Ma; e.g., Ehrmann and Mackensen, 1992;

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Zachos et al., 1994, 2001), but global climate warmed again until an optimum in the late early Miocene (17–15 Ma), the warmest interval of the Neogene. The cause of the early Miocene warming remains unclear, particularly since atmospheric CO₂ concentrations were probably not significantly higher than pre-industrial levels (Flower, 1999; Pagani et al., 1999a,b, 2005).

The early Miocene climate optimum was followed by rapid middle Miocene cooling, at a time when the influx of warm, relatively salty Tethyan Outflow Water (TOW, Wright et al., 1992; called Tethyan/Indian Saline Water [TISW] by Woodruff and Savin, 1989) into the north-western Indian Ocean has been argued to have ended, while the production of Northern Component Water (NCW), the Miocene equivalent of North Atlantic Deep Water, may have increased considerably (e.g., Woodruff and Savin, 1989; Wright et al., 1992; Wright and Miller, 1996; Roth et al., 2000; Lear et al., 2003). There is, however, no general agreement among workers on the reconstruction of Miocene ocean circulation in general, the volume of NCW and TOW in particular, and the cause of the rapid middle Miocene intensification of glaciation. Earlier papers (see above) generally sought the explanation in changes in deep-water circulation patterns, but more recent papers argue that orbitally driven changes in insolation were the main trigger (Abels et al., 2005; Holbourn et al., 2005). The latter authors argue that the orbital changes may have been reinforced by declining levels of atmospheric greenhouse gases, although records of such gases do not confirm this (Pagani et al., 1999a; Zachos et al., 2001), possibly because they are not available at high temporal resolution.

In the present Indian Ocean, the Asian summer (southwest) and winter (northeast) monsoons, characterized by a seasonally reversing wind system, have a profound effect on rainfall, runoff and vegetation in South Asia, as well as on the biota (e.g., Wang et al., 2005). The intense, wet, southwesterly winds of the summer monsoon cause extensive upwelling and high surface productivity in the Indian Ocean, particularly in the Arabian Sea (e.g., Schott and McCreary, 2001). Conversely, during the prevalence of dry northwesterly winds of the winter monsoon, surface productivity is relatively low (e.g., Schott and McCreary, 2001). The past climate in the Indian Ocean has been strongly influenced by the uplift of the Himalayan–Tibetan Plateau (Prell and Kutzbach, 1992; Molnar et al., 1993; Prell and Kutzbach, 1997; Niitsuma and Naidu, 2001). Various proxies (e.g., oceanic microfossils, land flora, stable isotopes, sedimentary records) have been used to suggest that the monsoon system initiated or increased

in intensity at 10–8 Ma as a consequence of this uplift (e.g., Quade et al., 1989; Kroon et al., 1991; Filipelli, 1997; An et al., 2001; Dettman et al., 2001). The main uplift may have occurred much earlier, however, i.e., before 15 Ma (middle Miocene; Mugnier and Huyghe, 2006) or even in the late Eocene (>35 Ma, Rowley and Currie, 2006), which would decouple the origin of the monsoons from the Himalayan uplift (see also Wang et al., 2005).

Gupta et al. (2004) also argued that the timing of, and interactions between, the elevation of the Himalayan–Tibetan Plateau, the onset of the monsoons and the changes in the Indian Ocean biota were poorly constrained, and suggested that the 10–8 Ma high-productivity event, which occurs over a large area including the Atlantic, Indian and Pacific Oceans (Dickens and Owen, 1999; Hermoyian and Owen, 2001), may have been the result of a strengthened wind system caused by global cooling and the increase in volume of the Antarctic ice sheets. This late Miocene interval of high oceanic productivity has now been documented to have occurred at many locations worldwide, such as the southeast (Diester-Haass et al., 2004) and northern Atlantic Ocean (Diester-Haass et al., 2005), the Pacific Ocean (Diester-Haass et al., in press) and the Tasman Sea (Tedford and Kelly, 2004; Diester-Haass et al., in press). Its causes thus are probably more global rather than directly linked to the development of the Indian Ocean monsoon. Possible effects on the global and Indian Ocean monsoonal circulation by the uplift of East Africa, argued to have become influential to atmospheric circulation by about 8 Ma, have not yet been evaluated (Sepulchre et al., 2006).

Benthic foraminifera dominate modern ocean floor meiobenthic and macrobenthic communities (e.g., Gooday et al., 1992; Gooday, 1999, 2003), and are the most abundant deep-sea organisms preserved in the fossil record. They are commonly used in paleoceanographic studies, but the environmental factors which influence their distribution and abundance on the ocean floor are complex and controversial (for a review, see e.g., Smart, 2002; Gooday, 2003; Jorissen et al., in press). The abundance and distribution of deep-sea benthic foraminifera are dominantly controlled by two, usually inversely related, parameters: the flux of organic matter (food) to the sea floor and the oxygen concentrations of bottom and pore waters (e.g., Lutze and Coulbourn, 1984; Gooday, 1988; Loubere, 1991; Gooday, 1993, 1994; Loubere, 1994; Jorissen et al., 1995; Loubere, 1996; Gooday and Rathburn, 1999; Loubere and Fariduddin, 1999a,b; Altenbach et al., 1999; Van der Zwaan et al., 1999; Gooday et al., 2000; Fontanier et al., 2002; Gooday, 2003; Fontanier et al.,

2005). As pointed out by Murray (2001), however, the situation is certainly more complex in many regions, and a combination of various parameters probably controls the composition of benthic foraminiferal faunas. These parameters include not only the amount and but also the nature of the supply of organic matter (food), oxygen concentration (including motion of redox fronts through the sediment, Fontanier et al., 2005), bathymetry, sediment type (grain size), and bottom water chemistry (e.g., carbonate undersaturation), hydrography (e.g., intensity of current flow) and hydrostatic pressure (e.g., Schnitker, 1994; Mackensen et al., 1995; Fontanier et al., 2002; Hayward et al., 2002; Jorissen et al., in press).

Benthic foraminiferal accumulation rates (BFARs) may provide a useful proxy for estimating the flux of organic matter to the ocean floor resulting from surface productivity (e.g., Herguera and Berger, 1991; Jorissen et al., in press). The use of this proxy in a quantitative sense, however, is not fully explored. The amount of organic matter arriving on the ocean floor is strongly depth dependent. Fluxes arriving at water depths of >1000 m represent only a few percent of the euphotic zone production (0.01–1.0%; e.g., Murray et al., 1996), and small differences in the transfer process of organic matter from surface to sea floor may strongly influence how much food arrives at the bottom. The process of transfer of organic matter from surface to sea floor is not well known in the present oceans. For instance, in the Pacific Ocean, there is a discrepancy between measured fluxes of sinking particulate organic matter and food demand (sediment community oxygen consumption), with the fauna apparently consuming more food than is supplied (Smith and Kaufmann, 1999; Smith et al., 2002). In addition, there is no linear correlation between productivity and flux below 2000 m depth at high productivities (>200 gC m⁻² year⁻¹) (Lampitt and Antia, 1997), and the correlation between BFAR and export productivity may not be linear in the presence of abundant opportunistically blooming, phytodetritus-exploiting species (e.g., Schmiedl and Mackensen, 1997).

In the present oceans, assemblages dominated by species of *Uvigerina*, *Bolivina*, *Bulimina* and *Melonis* generally occur in areas with a high, continuous flux of organic matter to the sea floor, possibly associated with lowered bottom water oxygen concentrations, such as observed in upwelling areas (e.g., Lutze and Coulbourn, 1984; Caralp, 1989; Sen Gupta and Machain-Castillo, 1993). In contrast, in well-oxygenated, generally oligotrophic open ocean settings, there are seasonal (and occasionally unpredictable) pulsed inputs of food in the form of phytodetritus (e.g., Thiel et al., 1990; Rice

et al., 1994; Smith et al., 1996; Kitazato et al., 2000), and opportunistic species of benthic foraminifera, notably *Epistominella exigua* and *Alabaminella weddellensis* (the so-called ‘phytodetritus species’), react dramatically to the seasonal arrival of phytodetritus by quickly colonizing and feeding on the detritus, and subsequently reproducing rapidly and building up large populations (e.g., Gooday, 1988, 1993, 1996; Sun et al., 2006). These ‘phytodetritus species’, and in particular *E. exigua*, have been used as proxies of pulsed organic matter fluxes in the geological past (e.g., Smart et al., 1994; Thomas et al., 1995; Ohkushi et al., 2000).

Benthic foraminifera occupy epifaunal and shallow to deep infaunal habitats (e.g., Corliss, 1985; Gooday, 1986; Jorissen, 1999). Microhabitat preferences have been related to calcareous test morphotypes, suggesting that a dominance of epifaunal morphotypes reflects higher oxygen concentrations and/or lower organic carbon contents, and dominance of infaunal morphotypes is indicative of lower oxygen concentrations and/or higher organic carbon contents (e.g., Corliss, 1985; Corliss and Chen, 1988; Roscoff and Corliss, 1991; Rathburn and Corliss, 1994). However, modern benthic foraminifera exhibit spatial and temporal variations in their microhabitats (Linke and Lutze, 1993; Bornmalm et al., 1997; Alve, 1999; Gross, 2000; Fontanier et al., 2002; Geslin et al., 2004) thus suggesting that there is no simple relationship between test morphology and microhabitat (e.g., Sun et al., 2006). Buzas et al. (1993) argued that such assignments for modern foraminifera may be accurate about 75% of the time. The categorisation of benthic foraminifera into epifaunal and infaunal groups may, however, provide a useful tool in paleoceanographic reconstructions, if used with caution (e.g., Thomas, 1990; Jorissen, 1999).

The distribution, biomass and diversity of Recent deep-sea benthic foraminifera in the northwest Indian Ocean (Arabian Sea) have been related to regional differences in bathymetry, bottom water oxygen concentrations and organic matter fluxes (Hermelin and Shimmield, 1990; Gupta, 1994; Jannink et al., 1998; Gooday et al., 1998, 2000; Kurbjeweit et al., 2000; Heinz and Hemleben, 2003). Numerous studies have been carried out on Neogene sequences from the northwestern Indian Ocean in an attempt to understand changes in water masses, productivity and biotic responses (e.g., Woodruff and Savin, 1989; Boersma, 1990; Boersma and Mikklesen, 1990; Kroon et al., 1991; Gupta, 1991; Hermelin, 1992; Gupta and Thomas, 1999; Kawagata et al., 2006). Most investigations have concentrated on understanding the evolution of the Asian Monsoon System in the Arabian Sea, and there have been few

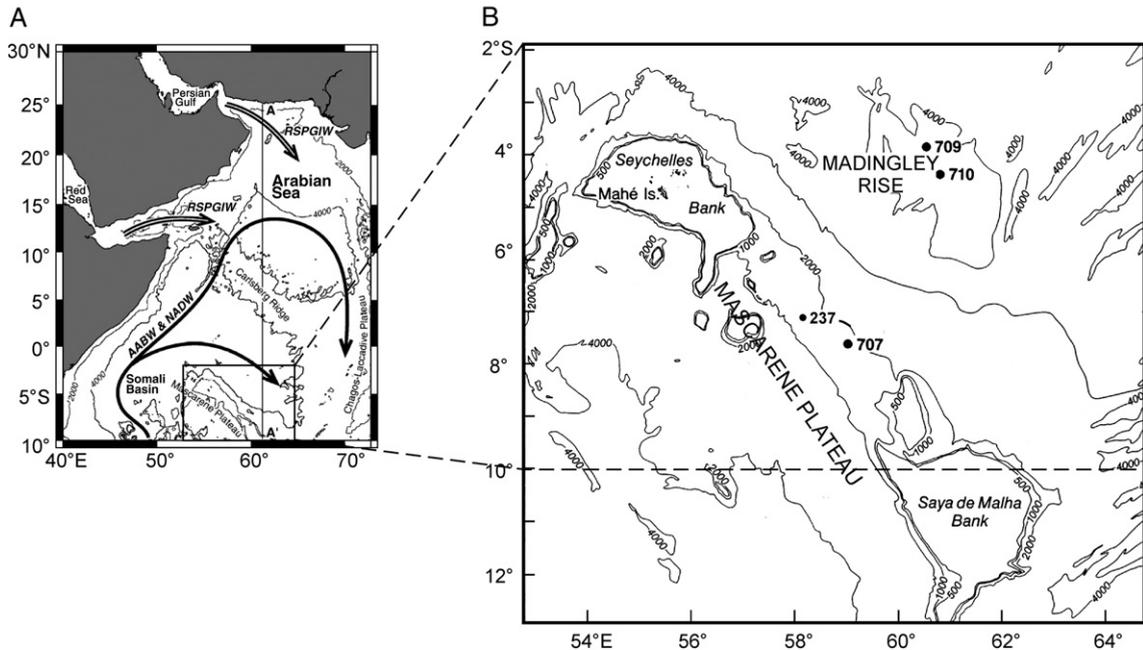


Fig. 1. (A) Location and bathymetric map of Sites 237, 707, 709 and 710 in the western equatorial Indian Ocean (bathymetry in meters), after Kawagata et al. (2006). Cross-section A–A': see Fig. 2. Arrows show present-day intermediate and bottom water flows (ABW=Antarctic Bottom Water; NADW=North Atlantic Deep Water; RSPGIW=Red Sea–Persian Gulf Intermediate Water). (B) Detailed location and bathymetric map of Sites 237, 707, 709 and 710 (after Fisher et al., 1974; Backman et al., 1988).

studies that have investigated the paleoceanographic history of the central western Indian Ocean.

Deep Sea Drilling Project (DSDP) Site 237 and Ocean Drilling Program (ODP) Sites 707, 709 and 710 represent a depth transect in the western equatorial Indian Ocean from the Mascarene Plateau to the Madingley Rise (1500 m to 3800 m) (Fig. 1). Upper Oligocene to upper Pliocene benthic foraminifera from Sites 707, and 709 and 710 were studied by Boersma (1990). Miocene benthic foraminiferal diversity fluctuations were studied by Smart and Murray (1995; Sites 709 and 710), Smart (1998; Sites 237, 707, 709 and 710). Aspects of early to middle Miocene deep-water paleoceanography (Sites 237, 707, 709 and 710) have been investigated by Smart and Ramsay (1995), Ramsay et al. (1998). We present new benthic foraminiferal faunal data from DSDP Sites 237 and ODP Sites 707, 709 and 710 and compare these with published stable isotope records. We aim to discuss long-term developments, and our data are at a relatively low resolution so that fluctuations at orbital timescales cannot be resolved.

2. Site locations, material and methods

DSDP Site (Hole) 237, ODP Site (Hole) 707A, ODP Site (Hole) 709C and ODP Site (Hole) 710A are located

in the western equatorial Indian Ocean (Fig. 1, Table 1). They lie on a depth transect from the Mascarene Plateau (Sites 237 and 707) to the Madingley Rise (Sites 709 and 710), a regional topographic high between the Carlsberg Ridge and the northern Mascarene Plateau. Sediments at these sites are mainly calcareous nannofossil oozes with good to moderate microfossil preservation. Generally complete Neogene sequences were drilled with good core recovery, although several unconformities (duration <0.5–1.0 Myr) may be present (Fisher et al., 1974; Backman et al., 1988). In addition, there is evidence of downslope transport and extensive reworking of older material over a thick part of the lower–middle Miocene section at Site 709 (Rio et al., 1990).

Table 1
Present location, present water depth and paleodepths of the study sites

Site	Present latitude	Present longitude	Present water depth (m)	Water depth at 5 Ma (m)	Water depth at 10 Ma (m)	Water depth at 15 Ma (m)
707	07°32.72'S	59°01.01'E	1552	1458	1362	1267
237	07°04.99'S	58°07.48'E	1623	1543	1463	1383
709	03°54.9'S	60°33.1'E	3040	2941	2846	2750
710	04°18.7'S	60°58.8'E	3824	3729	3624	3520

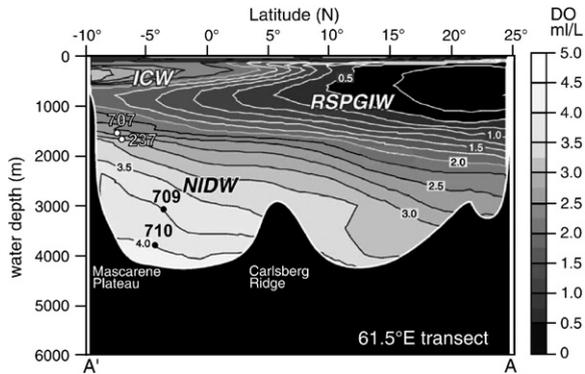


Fig. 2. Vertical profiles of dissolved oxygen content (DO ml/l) and water masses in the western Indian Ocean (modified after Kawagata et al. (2006)), with locations of Sites 237, 707, 709 and 710 (ICW=Indian Central Water; RSPGIW=Red Sea–Persian Gulf Intermediate Water; NIDW=North Indian Deep Water).

At present, surface water circulation of the northern and equatorial Indian Ocean is dominated by the seasonally changing monsoonal gyre system which is driven by the reversal of the prevailing wind field between the SW and NE monsoons. This reversing gyre is separated from the subtropical gyre by a hydrochemical front at $\sim 10^{\circ}\text{S}$ (Wyrki, 1973). To the north of this front, surface waters have high nutrient levels supporting high productivity, whereas to the south there are lower levels. The warm South Equatorial Current flows across the region between 5°S and 20°S (Tomczak and Godfrey, 1994).

A high-salinity, warm water mass (Red Sea–Persian Gulf Intermediate Water, RSPGIW), derived from the Red Sea and Persian Gulf, is a major component of intermediate waters in the northwestern Indian Ocean occupying depths above about 1000–1200 m (Fig. 2) (Tomczak and Godfrey, 1994; Kawagata et al., 2006). Below RSPGIW is North Indian Deep Water (NIDW), occupying depths from approximately 1500–2000 m to 4000 m (Wyrki, 1973; Tomczak and Godfrey, 1994; Kobayashi and Suga, 2006) [North Atlantic Deep Water (NADW) of Corliss, 1983] (Fig. 2). It has a high salinity and is created by mixing of Antarctic Bottom Water (AABW), Antarctic Intermediate Water (AAIW) and NADW with Red Sea–Persian Gulf outflow water to the north of the 10°S hydrochemical front (e.g., Schmitz, 1995) (Fig. 2). AABW, commonly referred to as Lower Circumpolar Deep Water (LCDW), is carbonate undersaturated and oxygen-rich (>4 ml O_2 l $^{-1}$) and flows northwards through the western basins and into the Mascarene Basin as a western boundary current below 3800 m, the lysocline depth (e.g., Wyrki, 1988; Reid, 2003) (Fig. 1). The study sites are presently within

NIDW, although an attenuated, distal part of RSPGIW influences the two shallower sites (237 and 707; Fig. 2).

Paleodepth reconstructions for each site (Table 1) were derived from subsidence curves (Ramsay et al., 1994). Depth relations between the sites have remained constant over the last 25 million years (Ramsay et al., 1994; Sykes et al., 1998), and all sites were situated within 10° south of the equator over the time interval studied (Sykes et al., 1998). Age models for Sites 707, 709 and 710 were constructed using the same datum levels as those selected by Peterson and Backman (1990), but assigning numerical ages according to the timescale of Lourens et al. (2004). Our age models for these sites thus are the same as those in Peterson and Backman (1990) in their interpretation of the biomagnetostratigraphy, with differences only in the update of the numerical ages of the datum levels. The age model for Site 237 was more difficult to develop and has considerably more uncertainty. The biostratigraphic data for this site (Roth, 1974; Vincent et al., 1974) were

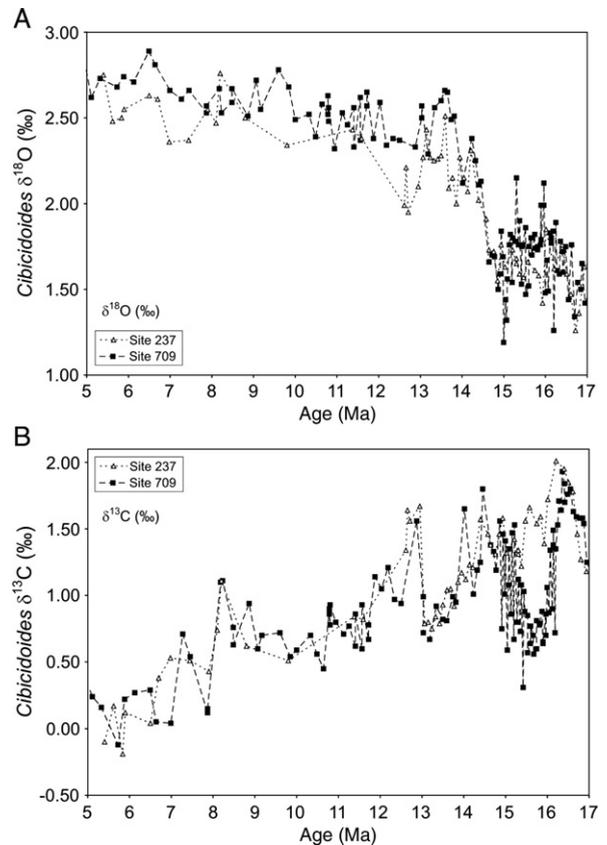


Fig. 3. (A) Benthic (*Cibicidoides* spp.) oxygen ($\delta^{18}\text{O}$) and (B) carbon ($\delta^{13}\text{C}$) isotope records (expressed in ‰ relative to PDB) for Site 237 (data from Woodruff and Savin, 1991) and Site 709 (data from Woodruff et al., 1990), with numerical ages after Lourens et al. (2004). See text for derivation of age model.

Table 2

Epifaunal species and species groups (after Corliss and Chen, 1988) used in this study

<i>Alabaminella weddellensis</i>	<i>Gavelinopsis</i> spp.	<i>Nuttallides umbonifera</i>
<i>Anomalinoidea</i> spp.	<i>Glabratella arctica</i>	<i>Oridorsalis umbonatus</i>
<i>Cibicidina</i> spp.	<i>Gyrogonoides</i> spp.	<i>Osangularia</i> spp.
<i>Cibicidoides</i> spp.	<i>Heronallenia lingulata</i>	<i>Planulina</i> spp.
<i>Epistominella exigua</i>	<i>Laticarinina pauperata</i>	<i>Pyrgo</i> spp.
<i>Eponides</i> spp.	<i>Lenticulina</i> spp.	<i>Quinqueloculina</i> spp.

collected at very low resolution (one sample per core, i.e., 9.6 m). We assigned numerical ages as in Lourens et al. (2004) to these datum levels to construct a preliminary age model, then assigned these preliminary ages to the isotope data for Site 237 (Woodruff and Savin, 1991). Finally, we optimized the fit between the oxygen and carbon isotope records of Site 237 and Site 709 so that differences between the isotope curves were minimal (Woodruff et al., 1990), and adjusted the numerical ages accordingly (Fig. 3A, B).

The carbon isotope curves for Sites 709 and 237 are significantly offset for the age interval between 16 and 15 Ma, with more negative values at Site 709. We think that this offset is an artifact caused by reworking; biostratigraphic data show major reworking (Peterson and Backman, 1990; Rio et al., 1990). The oxygen isotope values for that interval do not show an offset, and the combination of the major offset in carbon isotope offset and the lack of offset in oxygen isotope data can be explained by mixing of older material (from lower in the Miocene) according to the ages of reworked material indicated by the nannofossil record (Rio et al., 1990), as also noted by Woodruff and Savin (1991, Fig. 7). This older Miocene material would, from compilations of isotopic data (e.g., Zachos et al., 2001) as well as from the Site 709 record (Woodruff and Savin, 1991), have oxygen isotope values similar to these of 16–15 Myr old, combined with the lower carbon isotope values observed in the interval with common reworking.

Middle to upper Miocene (17–5 Ma) samples were collected at a resolution of approximately 0.5 Myr. Samples were washed through a 63- μ m sieve, dried and specimens of benthic foraminifera were picked from the >63 μ m size fraction to ensure that the paleoecologically important smaller sized specimens were included in the analysis (e.g., Schröder et al., 1987; Thomas et al., 1995). More than 250 specimens were picked from each sample and the relative abundance (percentage) and the numbers of benthic foraminifera per gram of total dry sediment were recorded. Benthic foraminiferal accumulation rates (BFARs, Herguera and Berger, 1991) were

calculated as follows: BFAR (number of specimens/cm² per kyr) = BF \times LSR \times DBD, where BF is the number of benthic foraminifera per gram of dry sediment, LSR is the linear sedimentation rate (cm/kyr), and DBD is the dry bulk density (g/cm³) of the sediment. Dry bulk density values were taken from the Initial Reports of the Ocean Drilling Program (Leg 115; Backman et al., 1988), and thus were not available for the specific levels of our samples. We used extrapolation from values for Sites 707, 709 and 710, which appears to be reasonable because there is little variability in the density values. For Site 237 only wet bulk values were published in the Site Reports (Fisher et al., 1974). We used the data published for Site 707 (similar in depth to Site 237) to derive an average ratio of wet to dry bulk density, and applied this ratio to estimate dry bulk density from the wet bulk density values provided for Site 237.

Our data are based on BFARs in the >63- μ m size fraction, whereas Herguera and Berger (1991) used the >150- μ m size fraction. An important part of the benthic foraminiferal assemblage consists of phytodetritus species, and these species are not well represented in the larger size fraction (e.g., Gooday, 1988). BFARs should always be cautiously interpreted because the values depend strongly on the estimates of sedimentation rates, which in turn depend upon reliable estimates of sediment numerical age. Relative abundance data are not dependent upon such estimates, and comparison of environmental information from BFAR and relative abundance data may assist in evaluating the reliability of the BFAR data.

We grouped calcareous benthic foraminifera into epifaunal and infaunal morphotypes following the categories of Corliss and Chen (1988). Epifaunal taxa include calcareous taxa with trochospiral and milioline test morphologies (Table 2). *E. exigua* and *Nuttallides umbonifera* are included in the epifauna category, but discussed separately because they make up a significant proportion of the epifauna (Table 3). High-productivity taxa are defined as the sum of the following infaunal taxa: *Bolivina* spp., *Bulimina* spp., *Melonis* spp. and *Uvigerina* spp. Percentages of elongate, cylindrical

Table 3

Mean percentages of *Epistominella exigua* and *Nuttallides umbonifera* comprising epifaunal species at the study sites

Site	<i>E. exigua</i> (percentage of epifauna)	<i>N. umbonifera</i> (percentage of epifauna)
707	19.2	25.7
237	21.0	28.4
709	25.8	16.7
710	21.0	28.8

benthic foraminifera (mainly stilostomellids and pleurostomellids, Hayward, 2002; Kawagata et al., 2005) were also recorded. For each sample, carbonate dissolution was recorded in terms of the percentages of planktic foraminiferal fragments (e.g., Le and Shackleton, 1992), calculated as the percentage of planktic foraminiferal test fragments relative to whole planktic foraminifera plus test fragments on counts of at least 300 planktic foraminiferal specimens.

3. Results

3.1. A new look at the benthic stable isotope records from Sites 237 and 709

The Miocene oxygen and carbon stable isotopes and paleoceanography from Sites 237 and 709 have been discussed by Woodruff and Savin (1989, 1991), Woodruff et al. (1990), Wright et al. (1992). These authors, however, did not plot the full time series for the

two sites on one timescale. The middle Miocene benthic foraminiferal oxygen isotope increase can be clearly recognized at both sites, with an age of 14.5–13.5 Ma at the resolution of these data. The major part of the isotope change has been dated at 13.91–13.84 Ma (Holbourn et al., 2005) or 13.82 ± 0.03 Ma (Abels et al., 2005) in high-resolution studies, within the range of our low-resolution estimate. The oxygen isotope data show great similarity between the deeper and shallower waters before the middle Miocene increase in values, with possibly somewhat increased temperature differences afterwards, with values at the shallower Site 237 generally lighter by about 0.5‰ (Fig. 3A, B).

The Miocene benthic foraminiferal carbon isotope record shows considerable variation on a 1–2 Myr timescale, so that errors in time correlation between sites may easily introduce apparent offsets, but the carbon isotope records for Sites 237 and 709 also show considerable similarity. The only interval where there is a strong offset between the two sites (16–15 Ma) is

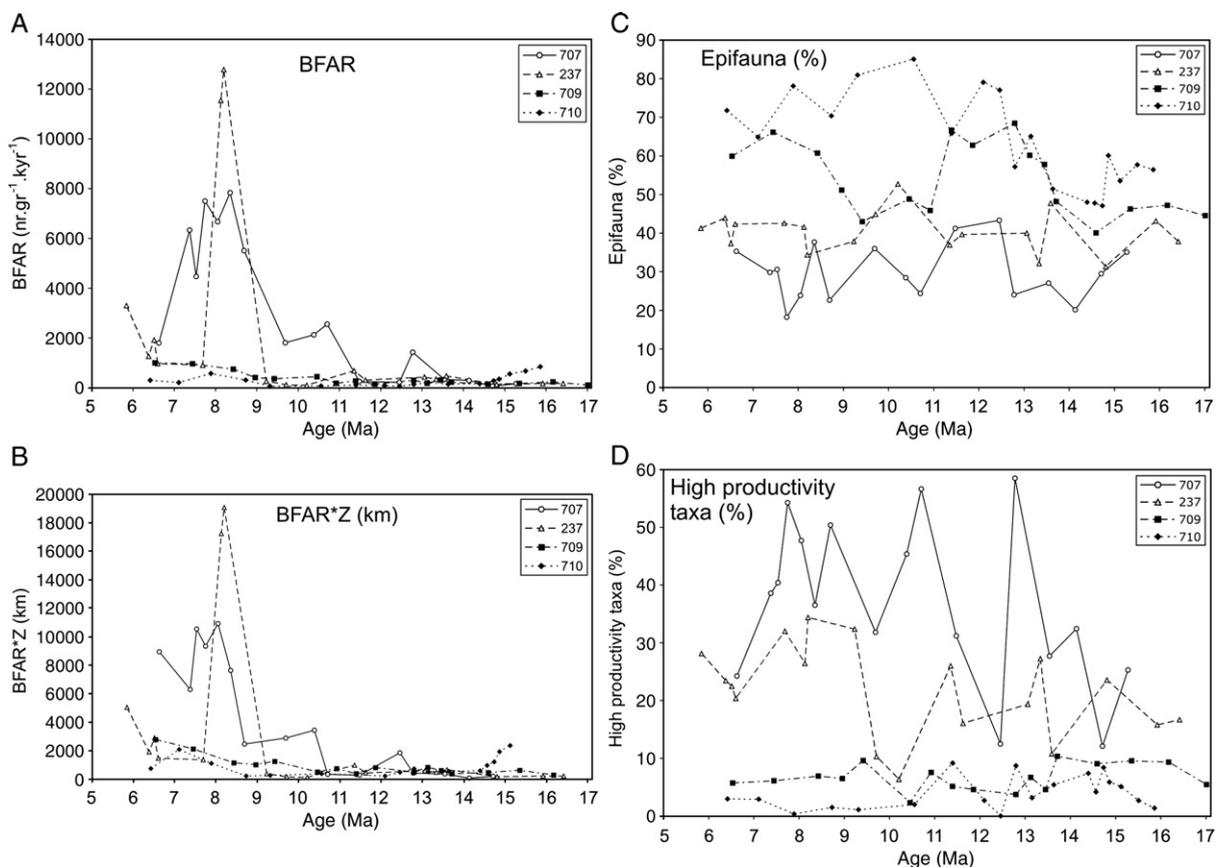


Fig. 4. Benthic foraminiferal faunal fluctuations at Sites 237, 707, 709 and 710: (A) benthic foraminiferal accumulation rates (BFARs) of total benthic foraminifera; (B) BFAR*Z index (Herguera and Berger, 1991), BFAR multiplied by paleowater depth (km); (C) percentage of epifauna; (D) percentage of high-productivity taxa (sum of *Bolivina* spp., *Bulimina* spp., *Melonis* spp. and *Uvigerina* spp.).

the interval with major reworking of older material, as discussed above. Benthic foraminifera with an age indicated by the nannofossil assemblages (Rio et al., 1990) would have the observed carbon and oxygen isotope signature (Woodruff and Savin, 1991).

The signature of the offset (lighter values at the deeper Site 709) could possibly be explained by the presence of an older water mass (derived from surface waters at a location further away from the western Indian Ocean) at depth than at shallower levels (Site 237). Generally, however, authors have argued that TISW water would be derived from sinking of waters at a location in the northernmost western Indian Ocean (e.g., close to the present Red Sea and Persian Gulf outlets), thus be younger and characterized by a heavier carbon isotope signature than southern-sourced waters (Woodruff and Savin, 1989). Such waters might be expected to also have significantly lighter oxygen isotope values (reflecting the higher temperatures, analogous to modern Mediterranean Outflow Water [MOW]; Woodruff and Savin, 1989; Seto, 1995). Our comparison of isotope data for Sites 237 and 709 shows no evidence for the presence of waters with a heavier

carbon isotope and lighter oxygen isotope signature at the deeper site in the earliest Miocene (before ~16 Ma).

We cannot exclude the possibility that the carbon isotope offset reflects the presence of a Tethyan-derived water mass (TOW) at the deeper Site 709 with a lighter rather than a heavier carbon isotope signature: such a water mass could be hypothesized to have been derived from waters sinking further west within Tethys, and thus further away from the Site 709. Such relatively older waters at Site 709 might be characterized by a light carbon isotope signature relative to southern-ocean derived intermediate waters at Site 237, especially if productivity in their region of origin was low. There is, however, no confirmation in the oxygen isotope data that such a warmer water mass is present, since values for both sites are not significantly different for the early Miocene. Of course, salinity effects and temperature effects might have cancelled, although in the modern MOW the temperature effect is dominant (Woodruff and Savin, 1989). Comparison of the records for Sites 237 and 709 thus suggests that these sites were not covered by significantly different water masses during the studied interval: if TISW was present at any one time,

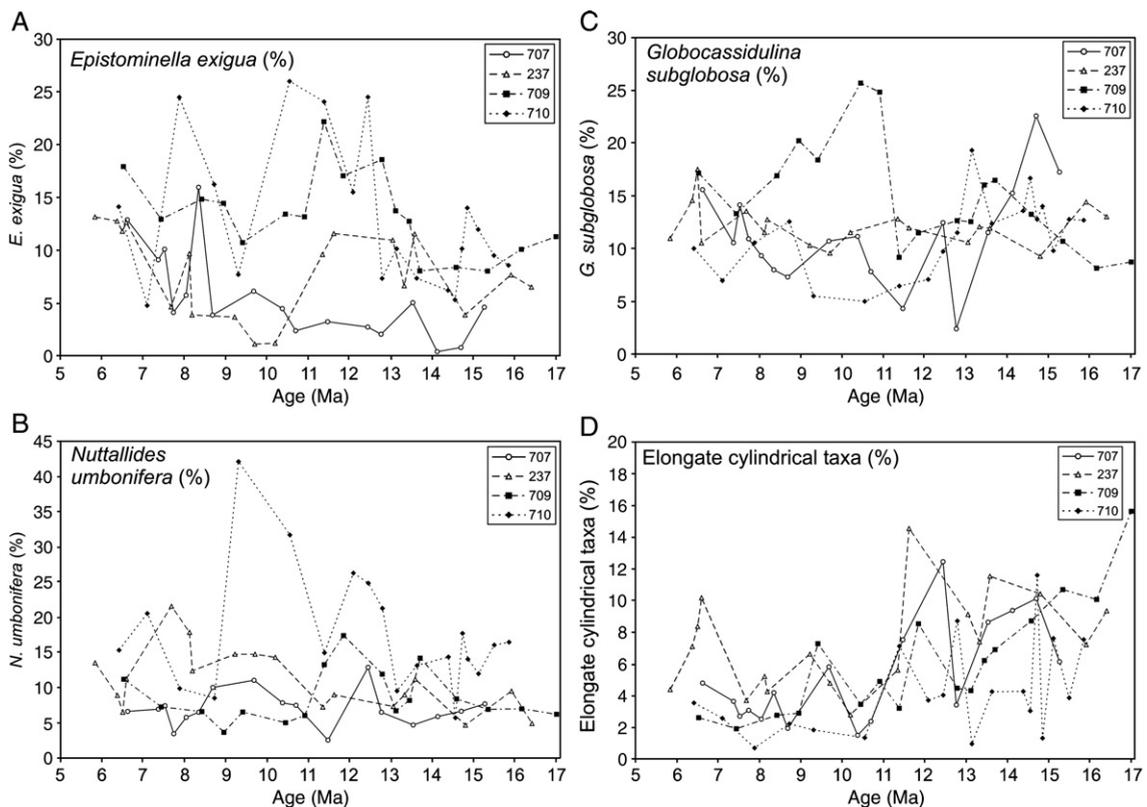


Fig. 5. Benthic foraminiferal faunal fluctuations at Sites 237, 707, 709 and 710: (A) percentage of *Epistominella exigua*; (B) percentage of *Nuttallides umbonifera*; (C) percentage of *Globocassidulina subglobosa*; (D) percentage of elongate, cylindrical taxa.

it must have covered both sites. Comparison of the oxygen isotope data from Sites 237 and 709 with the Zachos et al. (2001) compilation shows no significant difference in oxygen isotope values with the ‘average ocean’ values. Comparison of the carbon isotope values with those in the compilation show that values at both Sites 237 and 709 were lighter (indicating relatively old, not relatively young waters) between about 15 and 13 Ma, but also between 12 and 9 Ma and 8 and 6 Ma, thus no change at the time of the mid-Miocene increase in oxygen isotopic values. In view of the relatively low time resolution of available data as well as the lack of an orbitally tuned age model for the various sites, we argue that evidence for the existence of TOW in the western Indian Ocean during the Miocene is not convincing.

We constructed our age model for Site 237 minimizing the oxygen and carbon isotope offsets, so that the overall minor differences between the records for the two sites are (in part) a direct result of our methods. We argue, however, that the construction of this age model was constrained by the requirement that it should not violate any information in the magneto-biostratigraphic data set. The fact that it was possible to construct an age model, which at the same time minimizes the differences in the isotopic record and is in agreement with magneto-

biostratigraphy, implies in our opinion that there is no conclusive evidence that there are major primary (non-reworked) differences between the isotopic records.

3.2. Benthic foraminiferal accumulation rates (BFARs)

Benthic foraminiferal accumulation rates of total benthic foraminifera should be considered with caution because of the uncertainties in the age models, thus accumulation rates, for the sites. The BFARs show a general, although not pronounced, decrease with increasing water depth (as expected), and strong temporal fluctuations within sites (Fig. 4A). Values were generally low (<1000) before about 11 Ma, then increased strongly at the shallower sites, where values peaked between 9.5–7.5 Ma. The fact that the peak values are higher and apparently occur over a shorter time interval at Site 237 than at Site 707 may be a function of the age model because of the lack of high-resolution biostratigraphic data available for Site 237. At the two deeper sites there is a small and gradual increase in values starting at about 11 Ma.

Food flux to the sea floor is supposedly scaled with $1/Z$ (in which Z is water depth) (Herguera and Berger, 1991). If the only cause for differences between sites

Table 4
Site 707 correlation coefficients (r) and significance (p) values of species, species groups and other variables discussed in text

	BFAR	% Epifauna	%High productivity taxa	% <i>E. exigua</i>	% <i>N. umbonifera</i>	% <i>G. subglobosa</i>	%Elongate cylindrical taxa	%Planktic foram fragments
BFAR	r	–	0.555	0.535	–0.132	–0.304	–0.664	–0.279
	p	–	0.010	0.013	0.307	0.118	0.002	0.140
	n	–	17	17	17	17	17	17
%Epifauna	r	–0.310	–0.655	0.361	0.317	0.064	0.431	0.281
	p	0.113	–	0.077	0.107	0.403	0.042	0.138
	n	17	–	17	17	17	17	17
%High productivity taxa	r	0.555	–0.655	–	–0.033	–0.673	–0.837	–0.721
	p	0.010	0.002	–	0.449	0.185	0.002	0.000
	n	17	17	–	17	17	17	17
% <i>E. exigua</i>	r	0.535	0.361	–0.033	–	–0.018	–0.032	–0.260
	p	0.013	0.077	0.449	–	0.472	0.451	0.157
	n	17	17	17	–	17	17	17
% <i>N. umbonifera</i>	r	–0.132	0.317	–0.232	–0.018	–	0.135	0.136
	p	0.307	0.107	0.185	0.472	–	0.303	0.301
	n	17	17	17	17	–	17	17
% <i>G. subglobosa</i>	r	–0.304	0.064	–0.673	–0.032	0.135	–	0.439
	p	0.118	0.403	0.002	0.451	0.303	–	0.039
	n	17	17	17	17	17	–	17
%Elongate cylindrical taxa	r	–0.664	0.431	–0.837	–0.344	0.136	0.439	0.754
	p	0.002	0.042	0.000	0.088	0.301	0.039	–
	n	17	17	17	17	17	17	–
%Planktic foram fragments	r	–0.279	0.281	–0.721	–0.260	0.388	0.525	0.754
	p	0.140	0.138	0.001	0.157	0.062	0.015	0.000
	n	17	17	17	17	17	17	–

Significant values ($p \geq 0.01$) are given in bold.

Table 5
Site 237 correlation coefficients (r) and significance (p) values of species, species groups and other variables discussed in text

		BFAR	% Epifauna	%High productivity taxa	% <i>E. exigua</i>	% <i>N. umbonifera</i>	% <i>G. subglobosa</i>	%Elongate cylindrical taxa	%Planktic foram fragments
BFAR	r	–	–0.178	0.471	0.006	0.334	0.064	–0.351	0.046
	p	–	0.240	0.024	0.490	0.088	0.401	0.076	0.429
	n	–	18	18	18	18	18	18	18
%Epifauna	r	–0.178	–	–0.645	0.008	0.450	–0.004	–0.222	0.591
	p	0.240	–	0.002	0.487	0.030	0.493	0.188	0.005
	n	18	–	18	18	18	18	18	18
%High productivity taxa	r	0.471	–0.645	–	0.055	0.216	0.101	–0.284	–0.469
	p	0.024	0.002	–	0.414	0.195	0.345	0.126	0.025
	n	18	18	–	18	18	18	18	18
% <i>E. exigua</i>	r	0.006	0.008	0.055	–	–0.292	0.320	0.469	–0.440
	p	0.490	0.487	0.414	–	0.120	0.098	0.025	0.034
	n	18	18	18	–	18	18	18	18
% <i>N. umbonifera</i>	r	0.334	0.450	0.216	–0.292	–	–0.173	–0.595	0.428
	p	0.088	0.030	0.195	0.120	–	0.247	0.005	0.038
	n	18	18	18	18	–	18	18	18
% <i>G. subglobosa</i>	r	0.064	–0.004	0.101	0.320	–0.173	–	–0.044	–0.222
	p	0.401	0.493	0.345	0.098	0.247	–	0.431	0.188
	n	18	18	18	18	18	–	18	18
%Elongate cylindrical taxa	r	–0.351	–0.222	–0.284	0.469	–0.595	–0.044	–	–0.414
	p	0.076	0.188	0.126	0.025	0.005	0.431	–	0.044
	n	18	18	18	18	18	18	–	18
%Planktic foram fragments	r	0.046	0.591	–0.469	–0.440	0.428	–0.222	–0.414	–
	p	0.429	0.005	0.025	0.034	0.038	0.188	0.044	–
	n	18	18	18	18	18	18	18	–

Significant values ($p \geq 0.01$) are given in bold.

was the water depth, the value of BFAR* Z (i.e., BFAR multiplied by the paleodepth of the sites in kilometers; Herguera and Berger, 1991) would be equal for those sites. A plot of the index BFAR* Z , however, shows a similar offset as between BFAR values (Fig. 4B), so that the differences in BFARs between the sites cannot be explained by the depth difference.

3.3. Benthic foraminiferal relative abundances

The percentage of total epifaunal taxa generally increases with increasing water depth, but shows temporal fluctuations within sites (Fig. 4C), although less dramatic than those in the BFAR. At ~13 Ma, the percentage of epifauna at all sites increased, most pronouncedly at the two deeper sites (Sites 709 and 710). After ~10 Ma, the relative abundance of epifauna at the shallower water sites (Sites 707 and 237) returned to pre-13 Ma values. At Site 709, epifaunal values decreased to pre-13 Ma values after 11 Ma, but increased again from ~9 Ma.

The combined percentages of high-productivity taxa generally decrease with increasing water depth, as expected, and show a complex record of strong temporal fluctuations (Fig. 4D). Large fluctuations in abundance occurred particularly at Site 707 with peaks at 13 Ma,

11 Ma, and 9–8 Ma. Percentages of high-productivity taxa were lower at Site 237 than at Site 707, with an increase in values at about 13.5 Ma, followed by peak values between 10 and 8 Ma. At the deeper Sites 709 and 710 the percentages remain below 10% over the studied interval. There is no significant positive correlation between high-productivity taxa and BFARs at any of the sites (Tables 4–7).

The percentages of *E. exigua* fluctuate at all sites but are generally lower at the shallower Sites 707 and 237 (Fig. 5A). In the deeper Sites 709 and 710, values of *E. exigua* did not exceed 15% prior to ~14 Ma, and increased significantly between ~13.5 and 13 Ma, with highest but strongly fluctuating values between 13 and 10 Ma. All sites, except Site 707, record a period of high abundances of *E. exigua* between about 13 and 10 Ma.

The abundance of *N. umbonifera* (Fig. 5B) is highest at the deepest water Site 710 with peaks centered at 13–12 Ma and 11–10 Ma. At the other sites, percentages of *N. umbonifera* rarely exceed 20%.

The relative abundance of *Globocassidulina subglobosa* shows large fluctuations at all sites (Fig. 5C), but a large peak (~25%) occurred at Site 709 only, between 11 and 10 Ma. There is no overall simple relationship between the relative abundances of *N. umbonifera* and

Table 6

Site 709 correlation coefficients (r) and significance (p) values of species, species groups and other variables discussed in text

	BFAR	% Epifauna	%High productivity taxa	% <i>E. exigua</i>	% <i>N. umbonifera</i>	% <i>G. subglobosa</i>	%Elongate cylindrical taxa	%Planktic foram fragments	
BFAR	r	–	0.433	–0.205	0.291	–0.110	0.248	– 0.640	0.494
	p	–	0.041	0.215	0.129	0.338	0.168	0.003	0.022
	n	–	17	17	17	17	17	17	17
%Epifauna	r	0.433	–	– 0.581	0.793	0.456	–0.232	– 0.583	0.506
	p	0.041	–	0.007	0.000	0.033	0.186	0.007	0.019
	n	17	–	17	17	17	17	17	17
%High productivity taxa	r	–0.205	– 0.581	–	– 0.691	–0.105	–0.158	0.330	– 0.590
	p	0.215	0.007	–	0.001	0.344	0.272	0.098	0.006
	n	17	17	–	17	17	17	17	17
% <i>E. exigua</i>	r	0.291	0.793	– 0.691	–	0.392	–0.031	–0.545	0.653
	p	0.129	0.000	0.001	–	0.060	0.453	0.012	0.002
	n	17	17	17	–	17	17	17	17
% <i>N. umbonifera</i>	r	–0.110	0.456	–0.105	0.392	–	–0.367	0.010	0.090
	p	0.338	0.033	0.344	0.060	–	0.074	0.485	0.365
	n	17	17	17	17	–	17	17	17
% <i>G. subglobosa</i>	r	0.248	–0.232	–0.158	–0.031	–0.367	–	–0.503	0.441
	p	0.168	0.186	0.272	0.453	0.074	–	0.020	0.038
	n	17	17	17	17	17	–	17	17
%Elongate cylindrical taxa	r	– 0.640	– 0.583	0.330	–0.545	0.010	–0.503	–	– 0.630
	p	0.003	0.007	0.098	0.012	0.485	0.020	–	0.003
	n	17	17	17	17	17	17	–	17
%Planktic foram fragments	r	0.494	0.506	– 0.590	0.653	0.090	0.441	– 0.630	–
	p	0.022	0.019	0.006	0.002	0.365	0.038	0.003	–
	n	17	17	17	17	17	17	17	–

Significant values ($p \geq 0.01$) are given in bold.

G. subglobosa, although at the deepest Site 710 the percentage abundance of these species is negatively correlated (Table 7).

The percentage of elongate, cylindrical taxa (stilostomellids and pleurostomellids) show fluctuations at all sites (Fig. 5D), have slightly higher values at the two shallower sites, and overall decreased at 12–11 Ma from about 10% at the shallower two sites to less than 5%.

3.4. Planktic foraminiferal fragmentation

Temporal variations in percentages of planktic foraminiferal fragments within the sites are evident, with percentage fragmentation overall increasing with water depth, as expected (Fig. 6). At the two deepest sites, fragmentation was generally lower before about 13.5 Ma. At the shallowest Site 707 values fluctuated strongly before ~12 Ma, and at Site 237 peak values occurred between 11 and 9 Ma. There is a negative correlation between the percentage of planktic foraminiferal fragments and high-productivity taxa at all sites (Tables 4–7).

There is a positive correlation between the percentage abundance of *N. umbonifera* and percentage of planktic foraminiferal fragments at Site 237 (Table 5), but not at the other sites (Tables 4, 6 and 7). At Sites 707

and 709, the relative abundance of *G. subglobosa* is positively correlated with the percentage of planktic foraminiferal fragments (Tables 4 and 6), although there is no such relationship at Sites 237 and 710 (Tables 5 and 7).

4. Discussion

Benthic foraminiferal faunas document major changes in paleoenvironmental conditions during the middle to late Miocene (17–5 Ma) along a depth transect in the central western Indian Ocean across the Mascarene Plateau. Significant changes occurred from ~14 to 13 Ma onwards, but the commonly inconsistent events in benthic foraminiferal taxa at the different sites illustrate the complexity of factors controlling their distribution and abundance (e.g., Murray, 2001). Inconsistencies between information from BFARs and relative abundance data may at least in part reflect uncertainties in the age models, because the accumulation rate values depend strongly on estimates of sedimentation rate whereas relative abundance data are accumulation rate independent. Alternatively, the lack of significant positive correlation between the abundance of high-productivity taxa and BFARs at any of the sites (Tables 4–7), could suggest that either one or both

Table 7

Site 710 correlation coefficients (r) and significance (p) values of species, species groups and other variables discussed in text

		BFAR	% Epifauna	%High productivity taxa	% <i>E. exigua</i>	% <i>N. umbonifera</i>	% <i>G. subglobosa</i>	%Elongate cylindrical taxa	%Planktic foram fragments
BFAR	r	–	–0.265	–0.228	–0.105	–0.401	0.271	0.159	–0.183
	p	–	0.136	0.174	0.335	0.045	0.130	0.258	0.234
	n	–	19	19	19	19	19	19	18
%Epifauna	r	–0.265	–	–0.642	0.667	0.575	–0.619	–0.556	0.697
	p	0.136	–	0.002	0.001	0.005	0.002	0.007	0.001
	n	19	–	19	19	19	19	19	18
%High productivity taxa	r	–0.228	–0.642	–	–0.269	–0.232	0.155	0.624	–0.614
	p	0.174	0.002	–	0.133	0.169	0.263	0.002	0.003
	n	19	19	–	19	19	19	19	18
% <i>E. exigua</i>	r	–0.105	0.667	–0.269	–	0.123	–0.424	–0.209	0.303
	p	0.335	0.001	0.133	–	0.307	0.035	0.195	0.111
	n	19	19	19	–	19	19	19	18
% <i>N. umbonifera</i>	r	–0.401	0.575	–0.232	0.123	–	–0.718	–0.053	0.318
	p	0.045	0.005	0.169	0.307	–	0.000	0.415	0.099
	n	19	19	19	19	–	19	19	18
% <i>G. subglobosa</i>	r	0.271	–0.619	0.155	–0.424	–0.718	–	–0.006	–0.424
	p	0.130	0.002	0.263	0.035	0.000	–	0.491	0.040
	n	19	19	19	19	19	–	19	18
%Elongate cylindrical taxa	r	0.159	–0.556	0.624	–0.209	–0.053	–0.006	–	–0.546
	p	0.258	0.007	0.002	0.195	0.415	0.491	–	0.010
	n	19	19	19	19	19	19	–	18
%Planktic foram fragments	r	–0.183	0.697	–0.614	0.303	0.318	–0.424	–0.546	–
	p	0.234	0.001	0.003	0.111	0.099	0.040	0.010	–
	n	18	18	18	18	18	18	18	–

Significant values ($p \geq 0.01$) are given in bold.

of these proxies do not simply reflect food arrival at the sea floor, but incorporate information on such factors as the quality of organic matter (labile, refractory) or seasonality (e.g., Jorissen et al., in press). In order to explain the patterns of distribution of benthic foraminifera, consideration of a broad range of environmental factors is necessary and even within one region different environmental factors may reach critical thresholds

singly or in combination at different times and in different places.

There is no simple relationship between benthic foraminiferal abundance fluctuations and deep-water stable isotope values in the Miocene of the western equatorial Indian Ocean. Woodruff and Savin (1989, 1991), Woodruff et al. (1990), Wright et al. (1992) agreed that the carbon isotope data for shallow and deep sites indicated the presence of TOW (characterized by high $\delta^{13}\text{C}$ values) in the deep western equatorial Indian Ocean (depths of Sites 709, 710) during the early Miocene, and that this outflow ended during the middle Miocene time of cooling and establishment of the Antarctic ice sheet (~ 14.5 – 13.5 Ma in our timescale, Fig. 3A). Woodruff (1985) had argued for a similar circulation pattern using benthic foraminiferal assemblage data to indicate the presence of a relatively warm, low oxygen water mass in the western Indian Ocean in the early Miocene, which could have functioned in transporting heat to the Antarctic Continent, preventing the formation of a large East Antarctic Ice Sheet.

We now know, however, that an East Antarctic Ice sheet has been present since at least the earliest Oligocene, and expanded in the middle Miocene (e.g., Holbourn et al., 2005). In addition, climate modelling suggests that deep-ocean heat transport from the

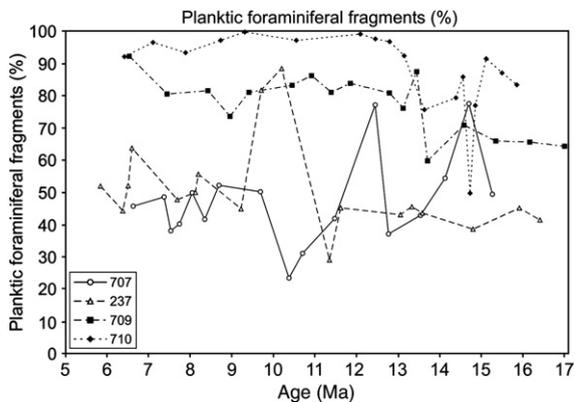


Fig. 6. Fluctuations in the percentage of planktic foraminiferal fragments (relative to whole planktic foraminifera) at Sites 237, 707, 709 and 710.

subtropics was not quantitatively sufficient to warm the Antarctic Continent (e.g., Huber and Sloan, 2001; Huber et al., 2004). We argue that it is not possible to decide unequivocally that TOW was present in the western equatorial Indian Ocean through the early Miocene using the presently available low-resolution data, and that earlier interpretations largely resulted from insufficiently precise age models and lack of time resolution. The possibility that TOW was present, however, cannot be refuted (as explained above), in part because of the lack of high-resolution data (including orbitally tuned age models).

Benthic faunal change started at ~14–13 Ma, with increased fragmentation at the deeper sites, increase in abundance of *N. umbonifera* at the deepest sites, increase in percentage of epifauna (especially %*E. exigua*) at the deeper sites, and increase in percentage of high-productivity taxa at the shallower sites. Further changes followed, centered at ~11 Ma, with even higher percentages of *N. umbonifera* at the deepest Site 710, increased abundance of *G. subglobosa* at the second deepest Site 709, a precipitous drop in percentage of elongate, cylindrical taxa, and in BFAR, especially at the shallowest two sites. These changes correspond in time and nature to global changes in benthic foraminiferal faunas that occurred towards the end of the middle Miocene (middle Miocene benthic foraminiferal turnover) (e.g., Woodruff, 1985; Thomas, 1985, 1986, 1987, 1992; Smart, 1998; Gupta et al., 2001; Thomas, in press). The decrease in abundance of the cylindrical species, observed worldwide (e.g., Thomas, in press) occurred later at our sites in the western Indian Ocean (~11 Ma, after the mid-Miocene increase in oxygen isotope values) than in the equatorial Pacific (Thomas and Vincent, 1987), where it clearly predated that increase. This overall decline (particularly 12–6 Ma) in the percentage of elongate, cylindrical taxa, whose abundances have been related to relatively low oxygen or high food for younger Neogene forms (Kawagata et al., 2005), appears to contrast with the overall evidence for increased productivity at our sites (see below). In addition, there is a significant negative correlation between the percentage of cylindrical taxa and BFAR and/or the percentage of high-productivity taxa at some of the sites. Possibly, these taxa required a more sustained and less seasonally variable food flux (see Thomas, in press), or were more competitive at higher deep-sea temperatures.

The BFAR and BFAR*Z data clearly indicate that overall supply of food to the benthics increased after about 11 Ma, but much more so at the two shallower sites on the Mascarene Plateau than at the deeper two

sites on Madingley Rise. The difference in depth cannot explain the difference in food supply to the bottom, suggesting that there were differences between primary productivity over the Mascarene Plateau and over Madingley Rise. We do not know what caused these differences in productivity, but increased vigor of deepwater circulation (AABW and NADW, Fig. 1) could have resulted in more vigorous vertical mixing and upwelling of nutrient waters against the Mascarene Plateau. The low, although slightly increasing values of BFAR throughout the studied section of the Miocene at the deepest sites might be in part due to increased dissolution, which is indicated by the high percentages of planktic foraminiferal fragments (e.g., Thunell, 1975; Le and Shackleton, 1992).

Increased surface productivity starting at about 13 Ma, intensifying at about 11 Ma, is reflected by the high abundance of the 'high-productivity taxa' (*Bolivina*, *Bulimina*, *Melonis* and *Uvigerina* species) at the shallowest Site 707, indicating high, continuous fluxes of organic matter to the sea floor (e.g., Lutze and Coulbourn, 1984; Caralp, 1989; Sen Gupta and Machain-Castillo, 1993). Interestingly, the correlation between BFAR and the abundance of 'high-productivity taxa' is not significant at any of the study sites, suggesting that there is no simple relationship between increased productivity and these proxies. We suggest that the increased overall productivity, especially during the high-productivity event (10–8 Ma) in the late Miocene, is expressed in very high BFAR values and high abundances of high-productivity taxa at the two shallower sites, where productivity may have been high for a larger part of the year, possibly due to more intense upwelling along the plateau. This coincides with increased mass accumulation rates of biogenic carbonate at Sites 707, 709 and 710 after 8 Ma (Peterson and Backman, 1990) and the worldwide 10–8 Ma high-productivity event (Dickens and Owen, 1999; Hermyoyian and Owen, 2001; Tedford and Kelly, 2004; Diester-Haass et al., 2004, 2005, in press). The overall increasing productivity at the deeper sites combined with an increase in seasonality may be expressed by the increasing abundance of *E. exigua*, an opportunistic species associated with seasonal inputs of phytodetritus to the ocean floor (e.g., Gooday, 1988, 1993; Smart et al., 1994; Gooday, 1996; Loubere, 1998). It is possible, therefore, that both seasonality and overall productivity increased in response to the initiation or intensification of the Asian Monsoon System at ~11 Ma, leading to a complex reaction of benthic foraminifera, in which high-productivity taxa dominated in some periods and at some locations (specifically where upwelling was high and

more pronounced), whereas at others the increased seasonality could not support taxa requiring a year-round high food supply. There is, however, no clear agreement on the timing of initiation of the Indian monsoon, with the early benthic foraminiferal faunal changes occurring before the commonly accepted date of 10–8 Ma (e.g., Kroon et al., 1991), but discussion is ongoing as to the possibility of an early monsoon initiation (Gupta et al., 2004; see also review in Wang et al., 2005).

The distribution of *N. umbonifera* has been associated with AABW or waters which are undersaturated with respect to calcium carbonate (Streeter, 1973; Bremer and Lohmann, 1982; Mackensen et al., 1995) and/or to areas of low productivity (Gooday, 1994; Smart and Gooday, 1997; Loubere and Fariduddin, 1999a). In the North Atlantic Ocean, the species occurs at lower latitudes, to the south of the strongly seasonal food input which leads to high abundance of phytodetritus species (Sun et al., 2006). Singh and Gupta (2004) proposed that in the late Oligocene to Miocene *N. umbonifera* was abundant in the southeastern Indian Ocean where other taxa could not grow optimally, either because of low food inputs or high carbonate corrosivity. Kurbjewit et al. (2000), however, argued that in the modern Arabian Sea, *N. rugosus* (= *N. umbonifera*) is not clearly related to either corrosive AABW or low productivity, but is significantly positively correlated with increased oxygen concentrations and sand content of the sediment. There is no simple relation at any of the study sites between the percentage abundance of *N. umbonifera* and percentage of planktic foraminiferal fragments which indicate increased dissolution (e.g., Thunell, 1975; Le and Shackleton, 1992). It is possible that the abundance increase in *N. umbonifera* at 11–9 Ma may be related to increased ventilation of bottom waters influencing the deepest Sites 709 and 710 in the western equatorial Indian Ocean.

G. subglobosa has a cosmopolitan distribution and its abundance has been related to a number of variables including various water masses (e.g., Corliss, 1979; Schnitker, 1980; Murray, 1988, 1991; Mackensen et al., 1995) and pulsed food inputs (Gooday, 1994; Gupta and Thomas, 2003). Corliss (1979) suggested that in the modern southeastern Indian Ocean, *Epistominella umbonifera* (= *N. umbonifera*) and *G. subglobosa* are associated with AABW, with *N. umbonifera* dominating where AABW is coldest (–0.2–0.4 °C) and *G. subglobosa* dominating where AABW is warmer (0.6–0.8 °C). In the modern eastern South Atlantic Ocean, the distribution of *G. subglobosa* has been associated with vigorous bottom currents and sandy sediments in elevated and oligotrophic areas (Schmiedl et al., 1997). At our sites, there is no simple relation-

ship between the abundances of *N. umbonifera* and *G. subglobosa*. We suggest that the high abundance of *G. subglobosa* at Site 709 may have been related to that same intensification of deep-water circulation that led to increased abundance of *N. umbonifera* at Site 710: the species co-vary in abundance in the northern Atlantic (Sun et al., 2006).

5. Conclusions

During the middle to late Miocene (17–5 Ma) significant and complex changes occurred in benthic foraminiferal faunas (>63 µm) in a depth transect from the western Indian Ocean (DSDP Site 237, ODP Sites 707, 709 and 710; water depth 1500–3800 m; paleodepths 1300–3500 m). A period of major faunal change occurred at ~14–13 Ma, possibly coeval with the middle Miocene increase in benthic foraminiferal oxygen isotope values, although this cannot be said with certainty because of the low time resolution of the available faunal and isotope data. Further faunal changes occurred at ~11–8 Ma, associated with assemblages indicating high productivity and food supply to the benthos. Inconsistent correlations between the relative abundances and accumulation rates of benthic foraminiferal taxa suggest that the factors controlling their distribution and abundance are not straightforward.

We argue that the benthic foraminiferal fauna changes can probably best be explained by the interplay between global cooling, intensification of deep-ocean circulation, thus sustained upwelling at some locations, but high seasonality of the food flux at other locations. The well-documented middle Miocene decrease in abundance of cylindrical species (e.g., Thomas, in press) did not occur at the same time in all oceans: it predated the middle Miocene isotope event in the equatorial Pacific, and post-dated it in the western Indian Ocean. The middle Pleistocene extinction of these forms likewise did not occur simultaneously in all oceans (Kawagata et al., 2005, 2006).

Benthic foraminiferal data have been interpreted as indicative of the presence of TOW in the western Indian Ocean (Woodruff, 1985), but modern faunal data suggest that the presence of a warm water mass with only slightly low oxygen values cannot be reliably inferred from faunal data (e.g., Gooday, 2003). We argue that neither the benthic foraminiferal assemblage data nor published stable isotope values (Fig. 3) indicate unequivocally that Tethyan Outflow Water (TOW) was present in the western equatorial Indian Ocean between 17 and 5 Ma at depths between 1300 and 3500 m. Evidence that the ending of such a relatively warm deep-

water flow was important in the increasing glaciation of the Antarctic continent thus is also not convincing.

Acknowledgements

We thank the Ocean Drilling Program for providing samples used in this study. The ODP was sponsored by the U.S. National Science Foundation (NSF) and participating countries (including the UK) under the management of the Joint Oceanographic Institutions (JOI), Inc. We are grateful to Ann Holbourn and an anonymous reviewer for their helpful comments. C.W.S. wishes to thank S. Jones for her continued support.

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