

**Response of eastern
Indian Ocean (ODP
Site 762B) benthic
foraminiferal assemblages
to the closure of the
Indonesian seaway**

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Abstract

Pliocene-Pleistocene deep sea benthic foraminifera from ODP Site 762B in the eastern Indian Ocean were examined to understand the tectonically/climatically induced palaeoceanographic changes. In addition to already published data on this site by Rai & Singh (2001), some more faunal parameters were considered in the present work. Characteristic benthic foraminiferal assemblages as well as more diverse fauna during the early Pliocene (before 3.5 Ma) reflected relatively oligotrophic and warm bottom water conditions. At the beginning of the late Pliocene (i.e. $\sim 3 \pm 0.5$ Ma) relative abundances of *Uvigerina proboscidea*, infaunal taxa and high productivity taxa increased, whereas faunal diversity showed a distinct decline, suggesting the development of pronounced upwelling resulting in higher surface water productivity. The strongly reduced inflow of warm and oligotrophic water masses as the South Equatorial Current (SEC) from the South Pacific to the eastern Indian Ocean due to the effective closure of the Indonesian seaway increased the surface water productivity. The closing of the Indonesian seaway during the late Pliocene was also responsible for the cessation of the warm, southward-flowing Leeuwin Current (LC) and the greater influence of the cold, deep

The complete text of the paper is available at <http://www.iopan.gda.pl/oceanologia/>

and northward-flowing Western Australian Current (WAC) in the eastern Indian Ocean.

1. Introduction

The Indonesian seaway is one of the critical zonal tropical seaways which largely influenced the global ocean circulation in the late Mesozoic, Paleogene and Neogene. The opening and closing of various seaways due to the drifting of continents significantly influenced climatic systems during most of the Cenozoic (Kennett et al. 1985). The long-term Cenozoic cooling trend is thought to have been initially stimulated by changes in the atmospheric circulation pattern resulting from the uplift of the Tibetan Plateau (Ruddiman et al. 1989, Raymao & Ruddiman 1992, Cerling et al. 1997). The change in the ocean circulation pattern following the opening of a continuous seaway around Antarctica at ~ 30 Ma was responsible for a fall in temperature in high latitudes (Toggweiler & Samuels 1995, Zachos et al. 2001). Significant changes in the circulation during the Pliocene as a result of several tectonic rearrangements in the tropics are believed to be the major causal mechanism for plunging the world into an ice age during the late Pliocene. The closure of the Indonesian seaway (Srinivasan & Sinha 1998, Cane & Molnar 2001, Gourelan et al. 2008) and the Panama seaway (Stehli & Webb (eds.) 1985, Burton et al. 1997, Bartoli et al. 2005) during the Pliocene affected the oceanic circulation, probably the deep thermohaline circulation. Deep sea records also provide ample evidence of changes in the thermohaline circulation (Burton et al. 1997). Rai & Singh (2001) have already published some of the data on faunal diversity and abundance to discuss the broad palaeoceanographic changes in this region. In the present paper several other faunal parameters are added for a better understanding of the response of the benthic foraminiferal distribution to the Indonesian seaway closure.

2. Indonesian seaway closure

In the course of the northward drift of Australia and Tasmania away from Antarctica, the Indonesian seaway between the Pacific and the Indian Ocean narrowed. Earlier studies suggested that the palaeoceanographic changes in the Indian Ocean, equatorial Pacific, South China Sea and Caribbean Sea were linked to the closure of the Indonesian and central American seaways during the Miocene and Pliocene (e.g. Keller 1985, Kennett et al. 1985, Haug & Tiedmann 1998, Srinivasan & Sinha 1998, Chaisson & Ravelo 2000, Haug et al. 2001, Jian et al. 2006). Through geological time, the position of the Indonesian seaway changed, as did the geometry of the inflow passages in relation to the tropical Pacific front, which significantly modified the

climatic role of the tropical Indian and Pacific Oceans, resulting in reduced atmospheric heat transport from the tropics to high latitudes (Nishimura 1992). The early evolution of the Indonesian seaway was characterized by tectonic restriction of the deep water and, by the early Miocene, the Indonesian seaway was already closed as a deep water pathway between the Pacific and Indian Oceans (Nishimura & Suparka 1997, Hall 2002, Kuhnt et al. 2002, 2004).

Hall (2002) suggested that the effective restriction in the Indonesian Throughflow (ITF) due to narrowing of the seaway could have occurred between 12 and 3 Ma. The remaining source of throughflow water shifted further north, resulting in a colder throughflow in the eastern Indian Ocean. A restriction of Indonesian Throughflow intensity at ~ 5 Ma was inferred from the significant expansion of the oxygen minimum zone in the central Indian Ocean (Dickens & Owen 1994). These authors concluded that the increased biological productivity was responsible for the expansion of the oxygen minimum zone in the central Indian Ocean as the warm oligotrophic Indonesian Throughflow water mass was strongly reduced. Srinivasan & Sinha (1998) also provided evidence for an early Pliocene restriction (at approximately 5 Ma) of the Indonesian seaway from a comparison of planktic foraminiferal species occurrences in the eastern Indian Ocean and tropical Pacific deep sea cores. Cane & Molnar (2001) suggested an even younger age (4–3 Ma) for the effective closure of the Indonesian seaway to restrict surface and thermocline water flow. They proposed that the emergence of the Indonesian Archipelago, in particular the rapid uplift of Halmahera dramatically reduced the Indonesian gateway. The past ocean circulation between the Pacific and Indian Oceans since the Miocene inferred from Nd isotopes (Gourlan et al. 2008) also supports the idea of the rapid closure of the Indonesian seaway around 4–3 Ma. Thus, various restriction events have been proposed for the middle Miocene, late Miocene, Pliocene and Pleistocene based on the circulation patterns in the equatorial Pacific Ocean and palaeoceanographic evidence from the Indian Ocean (Kuhnt et al. 2004). The final closure of the Indonesian seaway during Pliocene (~ 4 –3 Ma) (Cane & Molnar 2001) changed not only the physicochemical characteristics of the surface and deep water masses but also the circulation pattern in the Pacific and Indian Oceans. These oceanographic changes influenced the composition of the benthic and planktic foraminiferal assemblages. The aim of the present work is therefore mainly to understand the response of the eastern Indian Ocean benthic foraminiferal distribution to the oceanographic and climatic changes resulting from the closure of the Indonesian seaway.

3. Oceanographic setting

ODP Site 762B was drilled on the Exmouth Plateau off the coast of northwest Australia (lat. $19^{\circ}53.24'S$; long. $112^{\circ}15.24'E$; water depth – 1360 m) in the eastern Indian Ocean (Figure 1). This site is situated within the deep Oxygen Minimum Zone (Wyrтки 1971) below the tropical to subtropical transition zone ($20^{\circ}S$ to $15^{\circ}S$) (Bé & Hutson 1977). The South Equatorial Current (SEC), West Australian Current (WAC), Leeuwin Current (LC) and South Java Current (SJC) make up the major current system in this region. The intermingling of waters between the Indonesian passage and the equatorial Indian Ocean along with the Western Pacific Warm Pool (WPWP) and Indonesian Throughflow (ITF) largely controls the oceanographic conditions in the eastern Indian Ocean (Tomczak & Godfrey 2001). The warm, less saline WPWP is formed by the westward-flowing North and South Equatorial Currents, which are driven by the trade winds blowing westwards in the equatorial zone (Tomczak & Godfrey 2001).

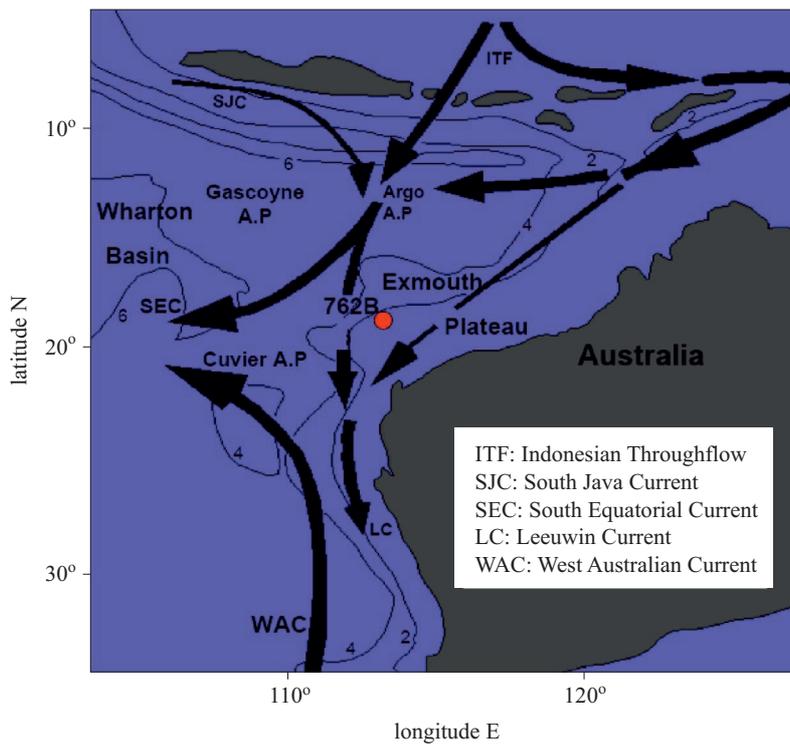


Figure 1. Oceanographic setting with major surface and subsurface currents and the location of ODP site 762B in the eastern Indian Ocean. Depth contours are in kilometres (from Rai & Singh 2001)

The less saline tropical Indonesian Throughflow (ITF) water originates in the WPWP and enters the Indian Ocean as the westward-flowing South Java and South Equatorial Currents. A part of the ITF starts flowing southwards along the western coast of Australia, around Cape Leeuwin and reaches as far as the Great Australian Bight as the Leeuwin Current (Cresswell & Golding 1980, Pearce 1991). Beneath the Leeuwin Current, high salinity waters are carried northwards by the cold Western Australian Current (WAC). This current is part of the major Southern Hemisphere subtropical gyre, moving anticlockwise in the Indian Ocean (Wells et al. 1994), which influences water masses to depths as great as 2000 m (Tchernia 1980).

The region of Exmouth off western Australia is geographically and topographically identical to the other eastern boundary regions. Therefore, the trade wind blowing equatorwards off western Australia would be expected to cause coastal upwelling in this region (Smith 1992). However, the ocean off western Australia behaves quite unlike other eastern boundary regions. There is no regular, continuous equatorward flow within 1000 km of the coast and no evidence of coastal upwelling. Coastal upwelling in this region is prevented or highly reduced by the warm, southward-flowing Leeuwin Current (LC), whose pressure gradient exceeds the off-shore Ekman transport (Smith 1992). However, there is strong indirect evidence for the development of zones of upwelling off the west coast of Australia during the glacial intervals (Wells et al. 1994). The examined ODP site is located in the region influenced by both the warm LC and the cold WAC. Thus, the fluctuations in the strength of these currents also affect the benthic foraminiferal distribution in this region.

4. Material and methods

The present study is based on 76 core samples from a 108.9 m thick section at ODP Site 762B in the eastern Indian Ocean. The core samples consist mainly of foraminifera-rich nannofossil ooze. Samples were wet sieved using $> 149 \mu\text{m}$ Tyler sieves. After drying, a micro-splitter was used to separate a representative portion of the $> 149 \mu\text{m}$ fraction estimated to contain about 300 specimens of benthic foraminifera. All the benthic foraminiferal specimens from the split samples were picked out and mounted on microfaunal assemblage slides for identification, counting and recording as percentages of the total assemblage.

In order to place greater stress on the dominant taxa, only species with a relative abundance of 2% or more in at least one sample and present in at least three samples were included in the Q-mode factor

analysis. The relative abundance of unilocular forms was not taken to perform factor analysis because their ecological preferences are not well known. Q-mode factor analysis was performed on a reduced data set of the 51 highest ranked species (Table 1) at this site using a commercially distributed statistical package (SPSS 9.0) to establish the correlation between benthic foraminiferal assemblages and environmental conditions. This method involves principal component analysis followed by VARIMAX rotation.

Table 1. Factor score of the 51 highest ranked benthic foraminiferal species

Species	Factor 1 (Score)	Factor 2 (Score)	Factor 3 (Score)	Factor 4 (Score)
<i>Textularia halkyardi</i>	-0.15216	-0.63443	-0.30301	-0.38262
<i>Textularia lythostrota</i>	-0.45838	0.27327	0.02575	0.38505
<i>Siphotextularia rolshauseni</i>	-0.08837	-0.47612	-0.39016	-0.61692
<i>Gaudryina solida</i>	-0.15912	0.06090	-0.39578	-0.13641
<i>Eggerella bradyi</i>	-0.12423	-0.17901	-0.15376	-0.36844
<i>Karreriella bradyi</i>	-0.05858	0.06209	0.05581	-0.53634
<i>Karreriella subrotundata</i>	-0.24283	-0.34087	-0.25757	-0.58419
<i>Quinqueloculina weaveri</i>	0.00830	-0.57965	-0.49758	0.58683
<i>Pyrgo depressa</i>	-0.04540	-0.64973	0.18576	-0.57137
<i>Biloculina murrhina</i>	0.18322	1.87601	0.80347	1.33200
<i>Sigmoilopsis schlumbergeri</i>	-0.03830	-0.36881	-0.40338	0.27334
<i>Stilostomella consobrina</i>	-0.19948	-0.32817	-0.64524	-0.42072
<i>Dentalina stimulea</i>	-0.21345	-0.08752	-0.27386	-0.72269
<i>Pyrrulinabiata</i>	-0.22098	-0.34072	-0.19266	-0.51518
<i>Ceratobulimina pacifica</i>	-0.18857	-0.66931	-0.52481	-0.07988
<i>Hoeglundina elegans</i>	0.02993	-1.09468	-0.50754	1.07408
<i>Sphaeroidina bulloides</i>	-0.28510	1.16177	0.17213	0.08877
<i>Bolivinitia pseudoplicata</i>	-0.18524	-0.45470	-0.48438	-0.48432
<i>Stilostomella insecta</i>	0.02619	-0.37172	1.51384	-1.94427
<i>Stilostomella lepidula</i>	0.03680	0.13876	1.04844	-0.34869
<i>Stilostomella subspinosa</i>	-0.19077	-0.29262	-0.67150	-0.66401
<i>Bulimina aculeata</i>	0.84703	-1.58322	-1.30350	3.68018
<i>Bulimina alazanensis</i>	-0.25909	-0.67929	2.42970	-1.43748
<i>Bulimina striata</i>	-0.13705	0.05134	-0.58723	-0.42586
<i>Uvigerina hispida</i>	-0.27039	-0.39071	-0.38054	-0.45688
<i>Uvigerina hispidocostata</i>	0.06417	-0.46887	-0.77503	1.20552
<i>Uvigerina peregrina</i>	0.16883	-0.51645	-1.31791	2.91038
<i>Uvigerina proboscidea</i>	7.36408	0.61809	0.17770	-0.95079
<i>Rectuvigerina royoii</i>	-0.39280	-0.14539	0.30600	-0.72713
<i>Epistomeinella exigua</i>	-0.17767	-0.40049	-0.26507	0.41110
<i>Gavelinopsis lobatulus</i>	-0.01913	-0.13436	-0.33510	0.17898

Table 1. *continued*

Species	Factor 1 (Score)	Factor 2 (Score)	Factor 3 (Score)	Factor 4 (Score)
<i>Laticarinina pauperata</i>	-0.12052	-0.04709	0.40509	0.25921
<i>Cibicides bradyi</i>	-0.01094	-0.06739	0.76767	0.01186
<i>Cibicides cicatricosus</i>	-0.32618	0.11668	-0.25616	-0.86732
<i>Cibicides kullenbergi</i>	-0.83217	2.68143	0.50142	0.72388
<i>Cibicides lobatulus</i>	-0.20642	3.21335	-1.86624	-1.13679
<i>Cibicides robertsonianus</i>	-0.16932	-0.43083	-0.50241	-0.39105
<i>Cibicides wuellerstorfi</i>	0.31158	-0.48430	4.22035	2.71820
<i>Pleurostomella acuta</i>	-0.18557	-0.54022	-0.31507	-0.65916
<i>Pleurostomella alternans</i>	-0.20265	-0.02984	-0.32390	-0.40525
<i>Cassidulina carinata</i>	0.07848	-0.47183	-0.93952	0.39024
<i>Ehrenbergina carinata</i>	-0.39405	-2.27210	3.39191	-0.56738
<i>Globocassidulina crassa</i>	-0.11566	-0.62185	-0.50275	-0.34962
<i>Globocassidulina subglobosa</i>	-0.04863	1.56339	0.64221	1.39683
<i>Globocassidulina murrhina</i>	-0.08821	-0.59340	-0.55339	-0.48237
<i>Astrononion umbilicatum</i>	0.00744	1.07458	-0.31196	-0.08547
<i>Pullenia bulloides</i>	0.05176	-0.53074	-0.08049	0.82753
<i>Pullenia osloensis</i>	-0.21103	-0.52174	-0.27963	-0.64061
<i>Pullenia quadriloba</i>	-0.20856	-0.18646	-0.07802	-0.16804
<i>Pullenia quenquiloba</i>	-0.24142	-0.33939	-0.39720	-0.06033
<i>Oridorsalis umbonatus</i>	-0.40567	2.83393	0.76614	1.14834
<i>Osangularia culter</i>	0.17417	0.26237	-0.19839	0.25217
<i>Osangularia</i> sp.	-0.15441	-0.21048	0.00321	-0.69691
<i>Gyroidinoides cibaoensis</i>	-0.39426	2.58576	1.54496	-0.02261
<i>Gyroidinoides nitidula</i>	-0.23637	0.10667	-0.65278	-0.26626
<i>Gyroidinoides orbicularis</i>	-0.15079	-0.47898	-0.51983	-0.49382
<i>Anomalina globulosa</i>	-0.27936	0.89426	-0.07424	-0.31122
<i>Anomalinoidea glabratus</i>	-0.20948	-0.00311	0.06969	-1.11614
1.5pt] <i>Melonis barleeaanum</i>	-0.05326	-0.55806	-0.51366	1.24007

The benthic foraminiferal quantitative data were used to calculate Hurlbert's diversity index, S_m (Hurlbert 1971). Hurlbert's diversity index is defined by the function

$$S_m = \sum_{i=1}^S [1 - C(N - N_i, m) / C(N, m)], \quad (1)$$

where S_m is the expected number of species in a random sub-sample of size m ($m \geq N$). In the present study $m=100$, which is well below the lowest number of specimens counted per sample. N is the number of specimens in the sample and S is the number of species in which

N specimens are distributed. N_i is the number of individuals in the i -th species, $\sum N_i = N$. $C(N - N_i, m) = (N - N_i)!/[m!(N - N_i - m)!]$ and $C(N, m) = N!/[m!(N - m)!]$ for $(N - N_i) \geq m$ and $N \geq m$ respectively, and zero for $(N - N_i) < m$ and $N < m$ respectively (Smith & Grassle 1977).

The percentages of shallow infaunal and other infaunal taxa were calculated following Wells et al. (1994), and the percentages of oxic and suboxic taxa were calculated following Kaiho (1994). We also compared the faunal diversity with some faunal abundance data. Benthic foraminifera were

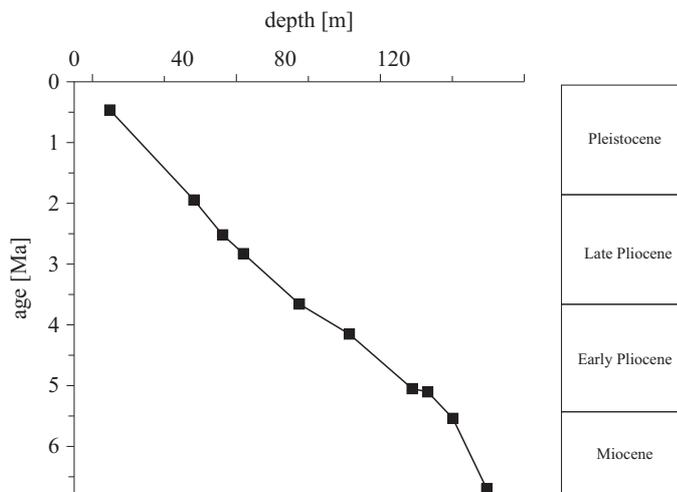


Figure 2. Age depth model of ODP site 762B (the details are given in Table 2)

Table 2. Calcareous nannofossil datums after Siesser et al. (1992). Astronomically tuned ages after Lourens et al. (2004) and Berggren et al (1995)

Depth [mbsf]	Age [Ma]	Event	Reference
4.9	0.46	<i>Pseudoemiliana lacunosa</i> LAD	Lourens et al. (2004)
28.2	1.95	<i>Discoaster brouweri</i> LAD	Lourens et al. (2004)
36.2	2.52	<i>Discoaster pentaradiatus</i> LAD	Lourens et al. (2004)
41.9	2.83	<i>Discoaster tamalis</i> LAD	Lourens et al. (2004)
57.4	3.66	<i>Sphenolithus</i> spp. LAD	Lourens et al. (2004)
71.4	4.15	<i>Discoaster asymmetricus</i> LAD	Berggren et al. (1995)
88.7	5.05	<i>Ceratolithus acutus</i> FAD	Lourens et al. (2004)
93.2	5.1	<i>Ceratolithus rugosus</i> FAD	Lourens et al. (2004)
100	5.54	<i>Discoaster quinqueramus</i> LAD	Lourens et al. (2004)
109.4	6.7	<i>Amaurolithus delicatus</i> FAD	Berggren et al. (1995)

grouped into percentages of total cylindrical elongate taxa (predominantly stilostomellids) following Hayward (2002) and Smart et al. (2007). High productivity taxa are explained as the sum of various infaunal taxa, i.e. *Bulimina* spp., *Melonis* spp., *Uvigerina* spp., *Ehrenbergina* spp., *Eggerella bradyi*, *Sphaeroidina bulloides* and *Pullenia bulloides* following Gooday (1994, 2003) and Loubere (1996).

The nannofossil datum levels, like those selected by Siesser et al. (1992), were used to construct the age model for Site 762B. But the numerical ages were reassigned according to the timescale of Berggren et al. (1995) and Lourens et al. (2004) (Figure 2). Our age model for this site is thus the same as that in Siesser et al. (1992) in their interpretation of the biomagnetostratigraphy, with differences only in the update of the numerical ages of datum levels (Table 2).

5. Results

Pliocene-Pleistocene deep sea benthic foraminifera show major fluctuations and long-term changes at ODP Site 762B (Figures 3 and 4). The most abundant species include *Uvigerina proboscidea*, *Cibicides lobatulus*, *Cibicides wuellerstorfi*, *Bulimina aculeata*, *Bulimina alazanensis*, *Stilostomella lepidula*, *Oridorsalis umbonatus* and *Gyroidinoides cibaoensis*. The benthic foraminiferal census records at Site 762B show a distinct change in species association and relative abundance at about 3 ± 0.5 Ma. On the basis of Q-mode factor analysis we recognized four distinct faunal assemblages at this site (Figure 5) and attempted to give their expected environmental preferences (Table 3).

Uvigerina proboscidea assemblage

U. proboscidea is the single dominant species of this assemblage, having a high positive score of factor 1. *U. proboscidea* is associated with the high organic carbon flux rates due to increased surface productivity and low oxygen levels resulting from organic matter oxidation (Gupta & Srinivasan 1992, Rai & Srinivasan 1994, Wells et al. 1994, Murgese & Deckker 2007). Thus, the *U. proboscidea* assemblage has been considered as an indicator of past periods of enhanced surface productivity (Table 3).

Cibicides lobatulus assemblage

Species of this assemblage have a distinct positive score of factor 2 comprising *C. lobatulus*, *O. umbonatus*, *Cibicides kullenbergi* and *G. cibaoensis*. *C. lobatulus* is an epiphyte species (Gaudant et al. 2010). *O. umbonatus* is a long-ranging species which lives in various environments

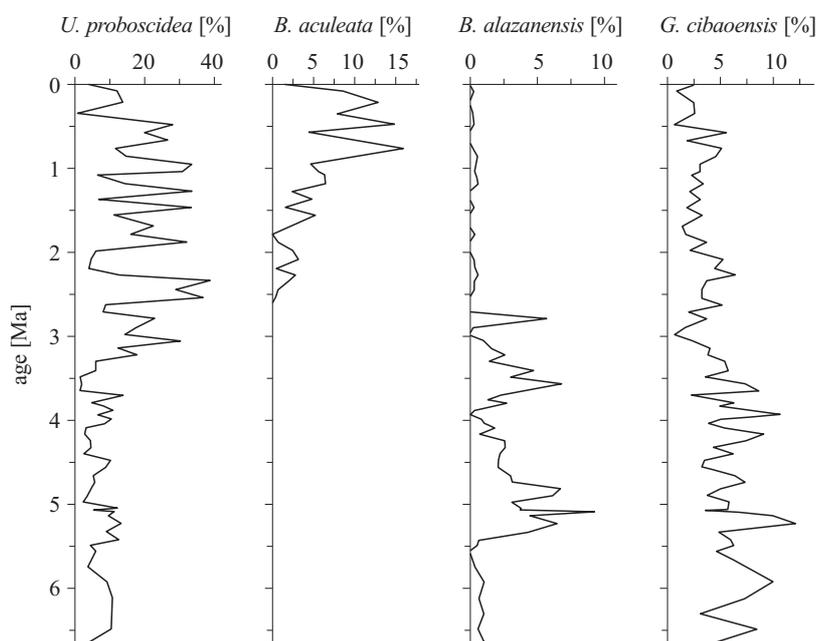


Figure 3. Time series plots of relative abundance of benthic foraminifera (*Uvigerina proboscidea* [%], after Rai & Singh 2001)

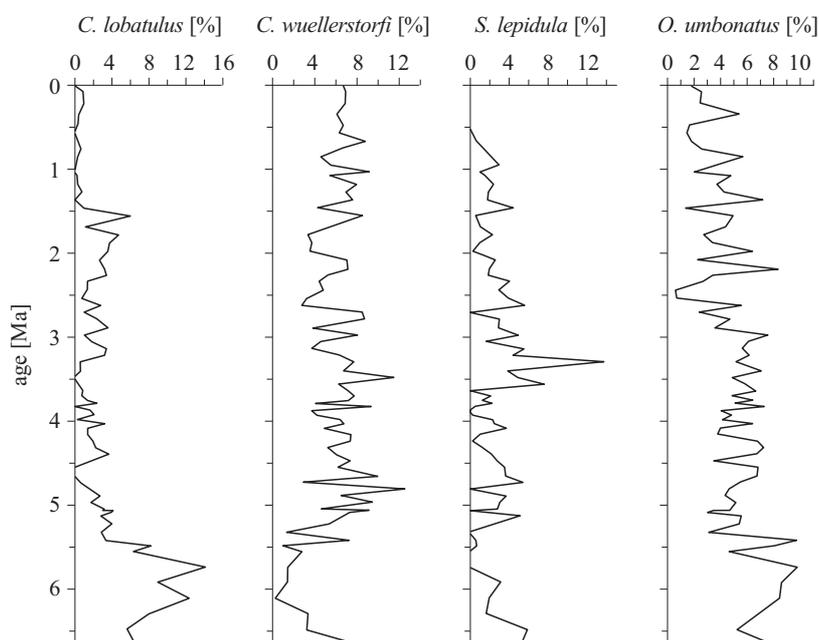


Figure 4. Time series plots of relative abundance of benthic foraminifera

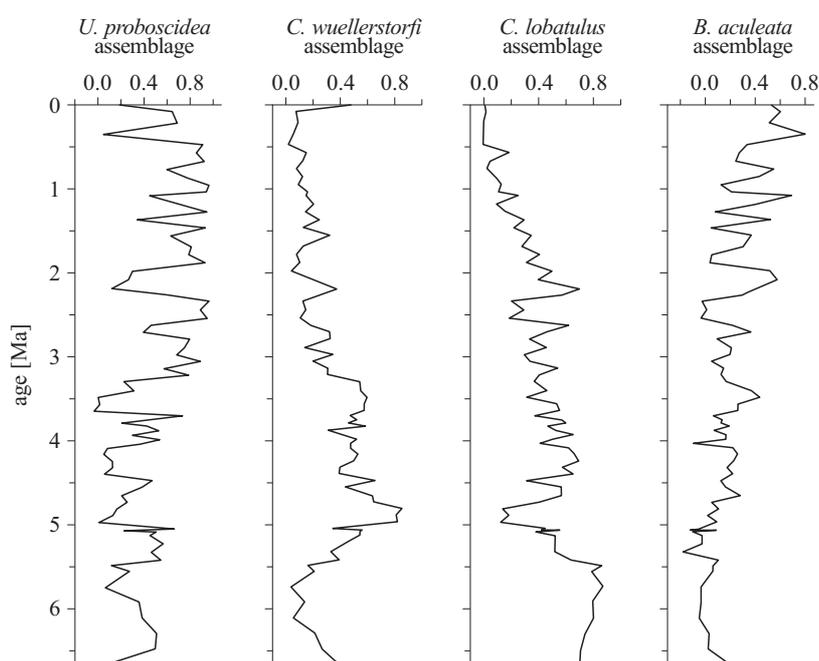


Figure 5. Time series plots of the loadings of the first four factors extracted on the basis of a Q-mode factor analysis of the benthic foraminiferal relative abundance data

Table 3. Benthic foraminiferal assemblages and their preferred environment

Q-mode assemblages	Variance [%]	Environmental preferences
<i>Uvigerina proboscidea</i> assemblage (factor 1)	29.82	increased surface productivity and continuous, high organic flux
<i>Cibicides lobatulus</i> assemblage (factor 2)	17.82	relatively warm, oxygenated bottom water with low organic flux
<i>Cibicides wuellerstorfi</i> assemblage (factor 3)	17.79	active currents, low-moderate organic flux, better oxygenation
<i>Bulimina aculeata</i> assemblage (factor 4)	9.7	oxygen poor and high organic flux

(Miao & Thunell 1993, Schmiedl & Mackensen 1997, Gupta & Thomas 1999). It is reported to reflect a well-oxygenated, low organic carbon environment (Mackensen et al. 1985, Miao & Thunell 1993). According to Rathburn & Corliss (1994) it can use limited amounts of food. *C. kullenbergi* prefers a deep-sea environment with a low organic carbon content below the low surface productivity regions (Burke et al. 1993, Nomura 1995).

The vertical distribution of *C. kullenbergi* is confined to the oxygen-rich and nutrient poor NADW (Schmiedl et al. 1997). *G. cibaoensis* is broadly distributed in the deep-sea environment with intermediate oxygen, and a variable temperature and food supply (De & Gupta 2010). This faunal assemblage is suggestive of an oxygenated deep-sea environment with a low organic flux (Table 3).

Cibicides wuellerstorfi assemblage

C. wuellerstorfi, *Ehrenbergina carinata*, *B. alazanensis*, and *G. cibaoensis* are the major species of this assemblage, with a high positive score of factor 3. As a suspension feeder and elevated epibiont, *C. wuellerstorfi* does not require high organic carbon levels and can withstand active bottom water currents (Linke & Lutze 1993, Gupta & Thomas 1999). *E. carinata* thrives in a warm deep sea with low oxygen and variable organic carbon levels (Nomura 1995, Gupta & Satpathy 2000). *E. carinata* is also reported from regions with an intermediate to high flux of organic matter and low oxygen conditions in the central Indian Ocean (Gupta et al. 2006). *B. alazanensis* is an infaunal species which thrives in a less well oxygenated deep sea with a high continuous food supply (Corliss & Chen 1988, Gupta & Thomas 1999). It is thus inferred that this faunal assemblage broadly reflects a low to intermediate flux of organic matter and oxygenated deep water with active currents (Table 3).

Bulimina aculeata assemblage

This assemblage is predominantly represented by *B. aculeata*, *U. peregrina* and *C. wuellerstorfi* with a relatively higher positive score of factor 4. *B. aculeata* thrives mainly in regions of relatively low to intermediate temperature with a low oxygen and high food supply (De & Gupta 2010). *U. peregrina* typically thrives in the deep sea with higher rates of organic carbon flux (Altenbach et al. 1999). This faunal assemblage is indicative of an oxygen-poor deep-sea environment with a high organic carbon flux (Table 3).

During most of the early Pliocene (prior to ~ 3.5 Ma) the low-food exploiting benthic foraminiferal assemblages (i.e. *C. lobatulus* and *C. wuellerstorfi* assemblages) developed significantly along with higher relative abundances of *C. lobatulus*, *C. wuellerstorfi*, *O. umbonatus* and *G. cibaoensis* (Figures 3 and 4). This time interval was also marked by a low percentage of total infaunal taxa and higher faunal diversity along with low abundances of taxa indicating higher surface water productivity and suboxic conditions (Figure 6). After ~ 3.5 Ma the typical

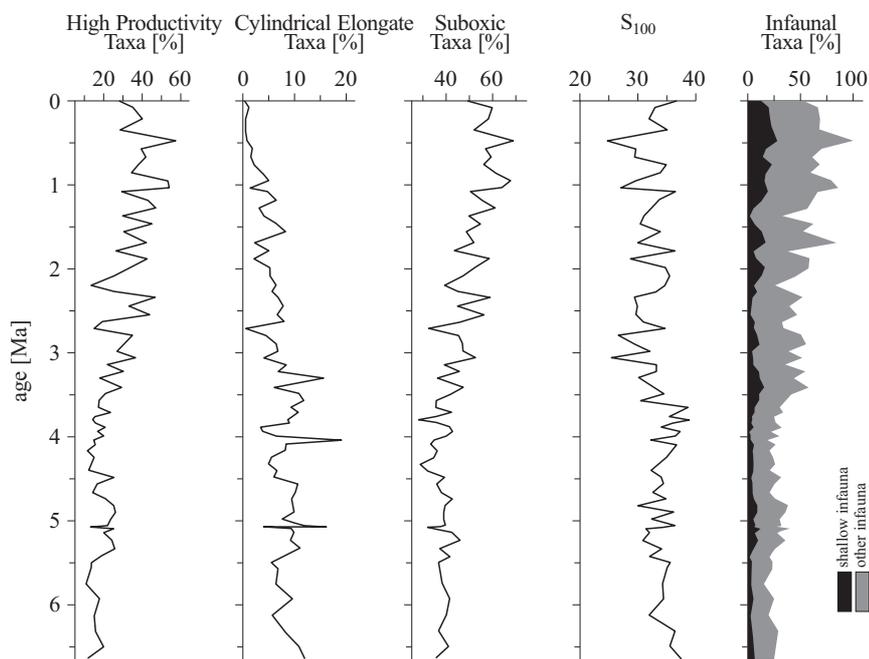


Figure 6. Time series plots of the percentage of high productivity taxa, cylindrical elongate taxa and suboxic taxa (present work); Hurlbert diversity indices (S_{100}) and percentage of total infaunal taxa (after Rai & Singh 2001)

high-food exploiting *U. proboscidea* assemblage started developing significantly, which was also marked by a regular increase in the relative abundance of *U. proboscidea*. At this time, the percentage of total infaunal taxa increased significantly, whereas species diversity showed a distinct decline (Figure 6). High-productivity taxa and suboxic taxa also started increasing their abundances at ~ 3.5 Ma and remained dominant during most of the late Pliocene and Pleistocene interval.

Most of the Pleistocene interval was characterized by the distinct development of the *B. aculeata* assemblage along with the *U. proboscidea* assemblage at this site (Figure 5). Interestingly, *B. aculeata* appeared at ~ 2.5 Ma (Figure 3), when *B. alazanensis* exhibited a sudden drop in its abundance, thereafter occurring sporadically during most of the late Pliocene and Pleistocene interval. Strong fluctuations in the relative abundance of *U. proboscidea* and the percentage of total infaunal taxa were observed during most of the Pleistocene. *S. lepidula* occurred more or less commonly during the Pliocene and early Pleistocene interval before disappearing in the middle Pleistocene, at a time coinciding with the absence of the *C. lobatulus* assemblage (~ 0.7 Ma) (Figure 4).

6. Discussion

Changes in the surface water productivity and climatically and/or tectonically induced ocean circulation may influence the deep-sea environment, causing variations in the benthic foraminiferal assemblages and species diversity (Thomas & Gooday 1996, Rai & Singh 2001, and others). Several recent studies have emphasized that variations in the organic carbon flux from the mixed layer due to the changing magnitude of surface water productivity play a vital role in the deep-sea benthic foraminiferal distribution pattern (Miao & Thunell 1993, Wells et al. 1994, Den Dulk et al. 1998, 2000, Rai et al. 2007). Thus, benthic foraminifera provide a useful tool for understanding the environmental variations in response to various palaeoclimatic and palaeoceanographic changes. In general, the early Pliocene interval (up to ~ 3.5 Ma) was characterized by the development of the *Cibicides lobatulus* and *Cibicides wuellerstorfi* assemblages, reflecting the low influx of organic carbon to the sea floor due to decreased rates of surface water productivity. This interval also corresponds with the increase in faunal diversity, and the relative abundance of *C. lobatulus* and *C. wuellerstorfi*, along with a low percentage of total infaunal taxa, high productivity taxa and suboxic taxa. Thus, during most of the early Pliocene the eastern Indian Ocean was characterized by relatively warm and stable bottom waters at bathyal depths with low organic carbon influx and better ventilation.

The early Pliocene was a period of prolonged global warming (Dowsett et al. 1996, Haywood et al. 2000) with relatively less developed Northern Hemisphere ice sheets and a higher CO₂ concentration (pCO₂) (Raymo et al. 1996). The low occurrence of the *C. lobatulus* assemblage along with the distinct occurrence of the *C. wuellerstorfi* assemblage at ~ 5 Ma could have been due to active bottom currents with a low-moderate organic flux in the deep sea. There is much evidence to suggest that up to the early Pliocene the Indonesian seaway was better open, which links the Indian Ocean and Pacific (Srinivasan & Sinha 1998, Cane & Molnar 2001) and most of the Indonesian Throughflow (ITF) was probably fed from warm south Pacific waters. Thus, up to ~ 3.5 Ma warm water from the southern Pacific should have been flowing into the Indian Ocean. Cane & Molnar (2001) proposed that a permanent El Niño-like condition during early the Pliocene increased the temperature at high latitudes and prevented the growth of Northern Hemisphere ice sheets. Thus, the inflow of warm southern Pacific waters in the eastern Indian Ocean during the early Pliocene increased the strength of the southward-flowing warm Leeuwin Current (LC), whose pressure gradient exceeds the offshore Ekman transport (Smith 1992), supplying tropical heat towards the pole. Ravelo et al. (2004) also explained that the lack of

a $\delta^{18}\text{O}$ gradient across the tropical Pacific during the early Pliocene warm period strongly supports a persistent El Niño-like condition increasing the eastern Indian Ocean temperature due to the intense Leeuwin Current. This, in turn, suppressed the intensity of the relatively deep and cold northward-flowing Western Australian Current (WAC), which may have been responsible for reduced upwelling and surface water productivity, as in modern times. Also during most of the early Pliocene (before ~ 3.5 Ma) the temperature of the deep ocean was significantly high and the thermocline was so deep that winds were unable to upwell deep, cold and nutrient-rich water to the surface (Fedorov & Philander 2000). The result was warmer sea surface temperatures in the Indian Ocean, elevated levels of evaporation and precipitation, and relatively wet climates in East Africa.

The late Pliocene (after ~ 3.5 Ma) was characterized by a distinct increase in the relative abundance of *Uvigerina proboscidea* (a well-known indicator of high surface water productivity; Gupta & Srinivasan 1992, Rai & Srinivasan 1994, Rai & Singh 2001, Rai et al. 2007, and others) and the significant development of high food-exploiting faunal assemblages (i.e. the *U. proboscidea* and *Bulimina aculeata* assemblages), along with a decrease in faunal diversity and higher percentages of total infaunal taxa. This was also a time of greater percentages of high-productivity taxa and suboxic taxa. The above faunal changes reflect the development of a strong upwelling-led high-productivity system at the beginning of the late Pliocene in the eastern Indian Ocean. Wells et al. (1994) also recorded identical benthic foraminiferal and isotopic signals in the eastern Indian Ocean during the penultimate glaciation and suggested an increase in surface water productivity due to the establishment of a zone of upwelling. The final closure of the Indonesian seaway during $\sim 4\text{--}3$ Ma changed the source of the Indonesian Throughflow (ITF) from the warm and saline south Pacific to the cooler and fresher north Pacific waters, which took a more westerly course. This, in turn, reduced the magnitude of the warm, southward-flowing Leeuwin Current and paved the way for the further northward flow of the cold Western Australian Current, which resulted in the marked shoaling of the thermocline in the eastern Indian Ocean. It was probably during this period that westerly equatorial winds also became stronger, which started to impinge on the west coast of Australia, and were accompanied by stronger tropical easterlies blowing off the Australian landmass (Venkatarathnam & Biscaye 1977). These stronger offshore winds are thought to have been responsible for the intense offshore Ekman transport, causing potential upwelling of cold and nutrient-rich water and the development of higher surface water productivity at low latitudes off the west coast of Australia in the eastern Indian Ocean. Karas et al. (2009) also attributed the gradual

freshening and related cooling ($\sim 4^{\circ}\text{C}$) of subsurface waters predominantly from ~ 3.5 to 2.95 Ma to the gradual constriction of the Indonesian seaway and the related switch in the source of subsurface ITF waters from the warm and saline south Pacific to the cooler and fresher north Pacific. At the same time, Lisiecki & Raymo (2005) recorded globally low values of benthic $\delta^{18}\text{O}$ with a small amplitude reflecting a low ice volume. The benthic Mg/Ca values do not suggest any distinct change in deep-sea temperatures either (Billups & Schrag 2002). Karas et al. (2009) argued that the significant cooling of Indian Ocean subsurface waters was not a result of the global cooling that intensified the Northern Hemisphere glaciations. Hence, we suggest that the tectonically driven changes in the regional current systems influenced the benthic foraminiferal assemblages at Site 762B in the eastern Indian Ocean.

The source of throughflow water further north due to the closure of Indonesian seaway and the resulting fall in SSTs in the eastern Indian Ocean would be responsible for reducing rainfall in eastern Africa. The increased gradient of sea surface temperature along with possible mountain building in New Guinea reduced the transport of heat from the tropics (the end of the Pliocene 'permanent El Niño') up to such a level as to cause global climatic cooling and the growth of ice sheets (Cane & Molnar 2001). These authors explained that changes in the Pacific Ocean dynamics resulting from the progressive closure of the Indonesian seaway triggered the transition from a permanent El Niño to the more La Niña-like climate of modern times. The new source of Pacific waters into the Indian Ocean, having changed from the southern warm thermocline to northern cold waters as a result of the northward drift of New Guinea across the equator (Rodgers et al. 2000), could have decreased SSTs in upwelling regions, which may in turn have caused a significant cooling of northern America through teleconnections and hence the initiations of the late Pliocene Northern Hemisphere glaciations. Earlier, Dickens & Owen (1994) inferred that the restriction of the warm and oligotrophic Indonesian Throughflow (ITF) from the Pacific to the Indian Ocean increased biological productivity, which was ultimately responsible for the expansion of the Oxygen Minimum Zone (OMZ) in the central Indian Ocean. They also suggested that before this closure warm water from the south Pacific was entering the Indian Ocean, increasing sea surface temperature and producing a rainier climate in eastern Africa.

The relative abundance of *U. proboscidea* and the percentage of total infaunal taxa increased considerably with much greater fluctuations during the Pleistocene. These faunal changes reflect prominent oscillations in the upwelling-led surface water productivity during the Pleistocene, possibly in

response to the episodic nature of the changing strength of the Leeuwin Current. The strength of the Leeuwin Current is largely dependent upon the behaviour of WPWP and Indonesian Throughflow waters (Godfrey & Weaver 1991) due to glacial and interglacial changes. Sinha et al. (2006) suggested that during glacial intervals the flow of the Leeuwin Current was substantially reduced or stopped altogether due to the reduction of WPWP and/or the lowering of the sea level, possibly as a result of intense cooling and ice formation. They also explained that the weakening of the southward-flowing Leeuwin Current resulted in a dominant equatorward wind-driven circulation, leading in turn to offshore Ekman transport and increased upwelling of cold, nutrient-rich water to the surface that enhanced surface water productivity in the eastern Indian Ocean.

B. aculeata, an infaunal species that prefers a higher organic matter input to the sea floor, appeared at ~ 2.5 Ma and increased its abundance thereafter. The increased magnitude of fluctuations in the relative abundances of *Bulimina aculeata*, species diversity and the percentage of total infaunal taxa during the Pleistocene reflects significant variations in the trophic level due to changing surface water productivity, possibly in response to glacial/interglacial changes. The relatively larger fluctuations in diversity values during the Pleistocene can be explained as a faunal response to the glacial and interglacial cycles with the varying eastern Asian monsoon regime (An 2000) and thus changing trophic conditions for the benthic foraminifera (Rai & Singh 2001). The disappearance of *Stilostomella lepidula* during the middle Pleistocene (~ 0.7 Ma) almost coincides with the so-called global '*Stilostomella*' extinction (Schönfeld 1996, Hayward 2001, 2002), whereas this species has also been recorded with rare and sporadic occurrences in the recent sediment of the Indian Ocean and other regions (see Rai & Singh 2004). The dominant occurrence of the *B. aculeata* assemblage during the last ~ 0.7 Ma suggests that the increase in upwelling and surface water productivity is possibly responsible for the sudden decline of oligotrophic *Stilostomella* and the almost complete absence of the *C. lobatulus* assemblage in the Indian Ocean. Gupta & Thomas (1999) also suggested that the decline and loss of this group was due mainly to intensified cooling combined with increased upwelling and surface water productivity, and in places the increased strength of tropical monsoons.

7. Conclusions

The closing of the Indonesian seaway was responsible for several palaeoceanographic changes in the eastern Indian Ocean. The final closure of the Indonesian seaway at about 4–3 Ma changed the source of the

Indonesian Throughflow (ITF) from warm south Pacific to cold north Pacific waters, which resulted in the breakup of permanent El Niño-like conditions. These changes reduced the warm thermocline water flow into the eastern Indian Ocean and also started the development of upwelling-led higher surface water productivity systems in this region. The flow of northern cold Pacific waters into the Indian Ocean may have lowered SSTs in upwelling regions, which caused the cooling of northern America through teleconnections and also initiated the late Pliocene glaciations in the Northern Hemisphere. The changing strength of the southward-flowing warm Leeuwin Current (LC) and the northward-flowing cold Western Australian Current (WAC) in response to glacial/interglacial cycles may have played an important role in the oceanographic setting of this area during the Pleistocene. These changes in the circulation pattern are the major controlling factors for upwelling, surface water productivity and deep sea ventilation, which ultimately influence the benthic foraminiferal distribution in this region. Thus, the Pleistocene glacial/interglacial cycles were responsible for the episodic nature of the flow of the Leeuwin Current in the eastern Indian Ocean, which resulted in marked fluctuations in surface water productivity.

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References

- Altenbach A. V., Pflaumann U., Schiebl R., Thies A., Timm S., Trauth M., 1999, *Scaling percentages and distributional patterns of benthic foraminifera with flux rates of organic carbon*, J. Foramin. Res., 29 (3), 173–185.
- An Z., 2000, *The history and variability of the East Asian paleomonsoon climate*, Quaternary Sci. Rev., 19 (1–5), 171–187, [http://dx.doi.org/10.1016/S0277-3791\(99\)00060-8](http://dx.doi.org/10.1016/S0277-3791(99)00060-8).
- Bartoli G., Sarnthein M., Weinelt M., Erlenkeuser H., Garbe-Schönberg D., Lea D. W., 2005, *Final closure of Panama and the onset of northern hemisphere glaciation*, Earth Planet. Sci. Lett., 237 (1–2), 33–44, <http://dx.doi.org/10.1016/j.epsl.2005.06.020>.
- Berggren W. A., Kent D. V., Swisher III C. C., Aubry M. P., 1995, *A revised Cenozoic geochronology and chronostratigraphy*, SEPM Spec. Publ., 54, 129–212.

- Bé A.W.H., Hutson W.H., 1977, *Ecology of planktonic foraminifera and biogeographic patterns of life and fossil assemblages in the Indian Ocean*, *Micropaleontology*, 23 (4), 369–414, <http://dx.doi.org/10.2307/1485406>.
- Billups K., Schrag D.P., 2002, *Paleotemperatures and ice volume of the past 27 Myr revisited with paired Mg/Ca and $^{18}\text{O}/^{16}\text{O}$ measurements on benthic foraminifera*, *Paleoceanography*, 17 (1), <http://dx.doi.org/10.1029/2000PA000567>.
- Burke S.C., Berger W.H., Coulbourn W.T., Vincent E., 1993, *Benthic foraminifera in box core ERDC 112, Ontong Java Plateau*, *J. Foramin. Res.*, 23 (1), 19–39, <http://dx.doi.org/10.2113/gsjfr.23.1.19>.
- Burton K.W., Ling H.F., O'niions R.K., 1997, *Closure of the Central American Isthmus and its effect on deep-water formation in the North Atlantic*, *Nature*, 386, 382–385, <http://dx.doi.org/10.1038/386382a0>.
- Cane M.A., Molnar P., 2001, *Closing of the Indonesian seaway as a precursor to east African aridification around 3–4 million years ago*, *Nature*, 411, 157–162, <http://dx.doi.org/10.1038/35075500>.
- Cerling T.E., Harris J.M., MacFadden B.J., Leakey M.G., Quade J., Eisenmann V., Ehleringer J.R., 1997, *Global vegetation change through the Miocene/Pliocene boundary*, *Nature*, 389, 153–158, <http://dx.doi.org/10.1038/38229>.
- Chaisson W., Ravelo A.C., 2000, *Pliocene development of the East-West hydrographic gradient in the equatorial Pacific*, *Paleoceanography*, 15, 497–505, <http://dx.doi.org/10.1029/1999PA000442>.
- Corliss B.H., Chen C., 1988, *Morphotype patterns of Norwegian Sea deep-sea benthic foraminifera and ecological implications*, *Geology*, 16 (8), 716–719, [http://dx.doi.org/10.1130/0091-7613\(1988\)016<0716:MPONSD>2.3.CO;2](http://dx.doi.org/10.1130/0091-7613(1988)016<0716:MPONSD>2.3.CO;2).
- Cresswell G.R., Golding T.J., 1980, *Observation of south-flowing current in the southeastern Indian Ocean*, *Deep-Sea Res. Pt. A*, 27 (6), 449–466, [http://dx.doi.org/10.1016/0198-0149\(80\)90055-2](http://dx.doi.org/10.1016/0198-0149(80)90055-2).
- De S., Gupta A.K., 2010, *Deep-sea faunal provinces and their inferred environments in the Indian Ocean based on distribution of Recent benthic foraminifera*, *Palaeogeogr. Palaeoclimatol.*, 291 (3–4), 429–442, <http://dx.doi.org/10.1016/j.palaeo.2010.03.012>.
- Den Dulk M., Reichart G.J., Heyst S.V., Zachariasse W.J., van der Zawn G.J., 2000, *Benthic foraminifera as proxies of organic matter flux and bottom water oxygenation? A case history from the northern Arabian Sea*, *Palaeogeogr. Palaeoclimatol.*, 161 (3–4), 337–359, [http://dx.doi.org/10.1016/S0031-0182\(00\)00074-2](http://dx.doi.org/10.1016/S0031-0182(00)00074-2).
- Den Dulk M., Reichart G.J., Memon G.M., Roelofs E.M.P., Zachariasse W.J., van der Zwaan G.J., 1998, *Benthic foraminiferal response to variations in the surface water productivity and oxygenation in the northern Arabian Sea*, *Mar. Micropaleontol.*, 35 (1–2), 43–66, [http://dx.doi.org/10.1016/S0377-8398\(98\)00015-2](http://dx.doi.org/10.1016/S0377-8398(98)00015-2).

- Dickens G.R., Owen R.M., 1994, *Late Miocene-early Pliocene manganese redirection in the central Indian Ocean: expansion of the intermediate water oxygen minimum zone*, *Paleoceanography*, 9, 169–181, <http://dx.doi.org/10.1029/93PA02699>.
- Dowsett H., Barron J., Poore R., 1996, *Middle Pliocene sea surface temperatures: a global reconstruction*, *Mar. Micropaleontol.*, 27 (1–4), 13–25, [http://dx.doi.org/10.1016/0377-8398\(95\)00050-X](http://dx.doi.org/10.1016/0377-8398(95)00050-X).
- Fedorov A., Philander G., 2000, *Is El Niño changing?*, *Science*, 288 (5473), 997–1002, <http://dx.doi.org/10.1126/science.288.5473.1997>.
- Gaudant J., Courme-Rault M., Saint-Martin S., 2010, *On the fossil fishes, diatoms, and foraminifera from Zanclean (Lower Pliocene) diatomitic sediments of Aegina Island (Greece): a stratigraphical and paleoenvironmental study*, *Palaeodiversity*, 3, 141–149.
- Gooday A.J., 1994, *The biology of deep-sea foraminifera; a review of some advances and their applications in Paleoceanography*, *Palaios*, 9 (1), 14–31, <http://dx.doi.org/10.2307/3515075>.
- Gooday A.J., 2003, *Benthic foraminifera (protista) as tools in deep-water palaeoceanography: environmental influences on faunal characteristics*, *Adv. Mar. Biol.*, 46, 1–90, [http://dx.doi.org/10.1016/S0065-2881\(03\)46002-1](http://dx.doi.org/10.1016/S0065-2881(03)46002-1).
- Godfrey J.S., Weaver A.J., 1991, *Is the Leeuwin Current driven by Pacific heating and winds?*, *Prog. Oceanogr.*, 27 (3–4), 225–272, [http://dx.doi.org/10.1016/0079-6611\(91\)90026-I](http://dx.doi.org/10.1016/0079-6611(91)90026-I).
- Gourlan A.T., Meynadier L., Allègre C.J., 2008, *Tectonically driven changes in the Indian Ocean circulation over the last 25 Ma: neodymium isotope evidence*, *Earth Planet. Sci. Lett.*, 26 (1–2), 353–364, <http://dx.doi.org/10.1016/j.epsl.2007.11.054>.
- Gupta A.K., Das M., Bhaskar K., 2006, *South Equatorial Current (SEC) driven changes at DSDP Site 237, Central Indian Ocean, during the Plio-Pleistocene: evidence from benthic foraminifera and stable isotopes*, *J. Asian Earth Sci.*, 28 (4–6), 276–290, <http://dx.doi.org/10.1016/j.jseaes.2005.10.006>.
- Gupta A.K., Satapathy S.K., 2000, *Latest Miocene-Pleistocene abyssal benthic foraminifera from west-central Indian Ocean DSDP Site 236: paleoceanographic and paleoclimatic inferences*, *J. Palaeontol. Soc. India*, 45, 33–48.
- Gupta A.K., Srinivasan M.S., 1992, *Uvigerina proboscidea abundances and paleoceanography of the northern Indian Ocean DSDP Site 214 during the Late Neogene*, *Mar. Micropaleontol.*, 19 (4), 355–367, [http://dx.doi.org/10.1016/0377-8398\(92\)90038-L](http://dx.doi.org/10.1016/0377-8398(92)90038-L).
- Gupta A.K., Thomas E., 1999, *Latest Miocene through Pleistocene paleoceanographic evolution of the northwestern Indian Ocean (DSDP Site 219): global and regional factors*, *Paleoceanography*, 14 (1), 62–73, <http://dx.doi.org/10.1029/1998PA900006>.
- Hall R., 2002, *Cenozoic geological and plate tectonic evolution of SE Asia and the SW Pacific: computer-based reconstructions, model and*

- animations*, J. Asian Earth Sci., 20 (4), 353–431, [http://dx.doi.org/10.1016/S1367-9120\(01\)00069-4](http://dx.doi.org/10.1016/S1367-9120(01)00069-4).
- Haug G.H., Tiedemann R., 1998, *Effect of the formation of the Isthmus of Panama on Atlantic Ocean thermohaline circulation*, Nature, 393, 673–676, <http://dx.doi.org/10.1038/31447>.
- Haug G.H., Tiedemann R., Zahn R., Ravelo A.C., 2001, *Role of Panama uplift on oceanic freshwater balance*, Geology, 29 (3), 207–210, [http://dx.doi.org/10.1130/0091-7613\(2001\)029<0207:ROPUOO>2.0.CO;2](http://dx.doi.org/10.1130/0091-7613(2001)029<0207:ROPUOO>2.0.CO;2).
- Hayward B.W., 2001, *Global deep-sea extinction during the Pleistocene ice ages*, Geology, 29 (7), 599–602, [http://dx.doi.org/10.1130/0091-7613\(2001\)029<0599:GDSEDT>2.0.CO;2](http://dx.doi.org/10.1130/0091-7613(2001)029<0599:GDSEDT>2.0.CO;2).
- Hayward B.W., 2002, *Late Pliocene to middle Pleistocene extinctions of deep sea benthic foraminifera ('Stilostomella Extinction') in the southwest Pacific*, J. Foramin. Res., 32 (3), 274–307, <http://dx.doi.org/10.2113/32.3.274>.
- Haywood A.M., Valdes P.J., Sellwood B.W., 2000, *Global scale palaeo-climate reconstruction of the middle Pliocene climate using the UKMO GCM, Initial results*, Global Planet. Change, 25 (3–4), 239–256, [http://dx.doi.org/10.1016/S0921-8181\(00\)00028-X](http://dx.doi.org/10.1016/S0921-8181(00)00028-X).
- Hurlbert S.H., 1971, *The nonconcept of species diversity: a critique and alternative parameters*, Ecology, 52 (4), 577–586, <http://dx.doi.org/10.2307/1934145>.
- Jian Z., Yu Y., Li B., Wang J., Zhang X., Zhou Z., 2006, *Phased evolution of the south-north hydrographic gradient in the South China Sea since the middle Miocene*, Palaeogeogr. Palaeoclimatol., 230 (3–4), 251–263, <http://dx.doi.org/10.1016/j.palaeo.2005.07.018>.
- Kaiho K., 1994, *Benthic foraminiferal dissolved oxygen index and dissolved oxygen levels in the modern ocean*, Geology, 22 (8), 719–722, [http://dx.doi.org/10.1130/0091-7613\(1994\)022<0719:BFDOIA>2.3.CO;2](http://dx.doi.org/10.1130/0091-7613(1994)022<0719:BFDOIA>2.3.CO;2).
- Karas C., Nürnberg D., Gupta A.K., Tiedemann R., Mohan K., Bickert T., 2009, *Mid-Pliocene climate change amplified by a switch in Indonesian subsurface throughflow*, Nat. Geosci., 2, 434–438, <http://dx.doi.org/10.1038/ngeo520>.
- Keller G., 1985, *Depth stratification of planktonic foraminifers in the Miocene Ocean*, Geol. Soc. Am. Mem., 163, 177–195.
- Kennett J.P., Keller G., Srinivasan M.S., 1985, *Miocene planktonic foraminiferal biogeography and paleoceanographic development of the Indo-pacific Region*, Geol. Soc. Am. Mem., 163, 197–236.
- Kuhnt W., Holbourn A., Hall R., Zuvella M., Käse R., 2004, *Neogene history of the Indonesian throughflow continent-ocean interactions with in East Asian marginal seas*, Geophys. Monogr. Ser., 149, <http://dx.doi.org/10.1029/149GM16>.
- Kuhnt W., Holbourn A., Zhao Q., 2002, *The early history of the South China Sea: evolution of Oligocene-Miocene deep water environments*, Rev. Micropaléontol., 45, 99–159.

- Linke P., Lutze G.F., 1993, *Microhabitat preferences of benthic foraminifera – a static concept or a dynamic adaptation to optimize food acquisition?*, Mar. Micropaleontol., 20 (3–4), 215–234, [http://dx.doi.org/10.1016/0377-8398\(93\)90034-U](http://dx.doi.org/10.1016/0377-8398(93)90034-U).
- Lisiecki L.E., Raymo M.E., 2005, *A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}O$ records*, Paleoceanography, 20 (PA1003), <http://dx.doi.org/10.1029/2004PA001071>.
- Loubere P., 1996, *The surface ocean productivity and bottom water oxygenation signals in deep water benthic foraminiferal assemblages*, Mar. Micropaleontol., 28 (3–4), 247–261, [http://dx.doi.org/10.1016/0377-8398\(96\)00004-7](http://dx.doi.org/10.1016/0377-8398(96)00004-7).
- Lourens L.J., Hilgen F.J., Laskar J., Shackleton N.J., Wilson D., 2004, *The Neogene period*, [in:] *A geologic time scale*, F. M. Gradstein, J. G. Ogg & A. G. Smith (eds.), Cambridge Univ. Press., Cambridge, 409–440.
- Mackensen A., Sejrup H.P., Janse E., 1985, *Living benthic foraminifera off Norway*, Mar. Micropaleontol., 9 (4), 275–306, [http://dx.doi.org/10.1016/0377-8398\(85\)90001-5](http://dx.doi.org/10.1016/0377-8398(85)90001-5).
- Miao Q., Thunell R., 1993, *Recent deep-sea benthic foraminiferal distributions in the South China and Sulu Seas*, Mar. Micropaleontol., 22 (1–2), 1–32, [http://dx.doi.org/10.1016/0377-8398\(93\)90002-F](http://dx.doi.org/10.1016/0377-8398(93)90002-F).
- Murgese S.D., Deckker P.D., 2007, *The late Quaternary evolution of water masses in the eastern Indian Ocean between Australia and Indonesia, based on benthic foraminifera faunal and carbon isotope analyses*, Mar. Micropaleontol., 247 (3–4), 382–401, <http://dx.doi.org/10.1016/j.palaeo.2006.11.002>.
- Nishimura S., 1992, *Tectonic approach to changes in surface-water circulation between the tropical Pacific and Indian Oceans*, [in:] *Pacific Neogene-environment, evolution and events*, R. Tsuchi & J.C. Ingle Jr. (eds.), Univ. Tokyo Press, 157–167.
- Nishimura S., Suparka S., 1997, *Tectonic approach to the Neogene evolution of the Pacific Indian Ocean seaways*, Tectonophysics, 281 (1–2), 1–16, [http://dx.doi.org/10.1016/S0040-1951\(97\)00155-8](http://dx.doi.org/10.1016/S0040-1951(97)00155-8).
- Nomura R., 1995, *Paleogene to Neogene deep-sea paleoceanography in the eastern Indian Ocean: benthic foraminifera from ODP Sites 747, 757 and 758*, Micropaleontology, 41 (3), 251–290, <http://dx.doi.org/10.2307/1485862>.
- Pearce A.F., 1991, *Eastern boundary currents of the southern hemisphere*, J. Roy. Soc. West. Aust., 74, 35–45.
- Rai A.K., Singh V.B., 2001, *Late Neogene deep-sea benthic foraminifer at ODP site 762B, eastern Indian Ocean: diversity trends and Paleoceanography*, Palaeogeogr. Palaeoclimatol., 173 (1–2), 1–8, [http://dx.doi.org/10.1016/S0031-0182\(01\)00299-1](http://dx.doi.org/10.1016/S0031-0182(01)00299-1).
- Rai A.K., Singh V.B., 2004, *Late Neogene deep sea benthic foraminiferal biostratigraphy of ODP sites 762B and 763A, (Exmouth Plateau) eastern Indian Ocean*, J. Geol. Soc. India, 63, 415–429.

- Rai A. K., Srinivasan M. S., 1994, *Pleistocene oceanographic changes indicated by deep sea benthic foraminifera in the northern Indian Ocean*, Proceedings of the Indian Academy of Sciences (Earth and Planetary Sciences), 103 (4), 499–517.
- Rai A. K., Srinivasan M. S., Maurya A. S., 2007, *Influence of antarctic bottom water (AABW) and monsoonal activity in the central Indian Ocean over past 5 million years: benthic foraminiferal record at DSDP site 238*, J. Geol. Soc. India, 69 (2), 222–231.
- Rathburn A. E., Corliss B. H., 1994, *The ecology of living (stained) deep-sea benthic foraminifera from the Sulu Sea*, Paleooceanography, 9 (1), 87–150, <http://dx.doi.org/10.1029/93PA02327>.
- Ravelo A. C., Andreasen D. H., Lyle M., Lyle A. O., Wara M. W., 2004, *Regional climate shifts caused by gradual global cooling in the Pliocene epoch*, Nature, 429, 263–267, <http://dx.doi.org/10.1038/nature02567>.
- Raymo M. E., Grant B., Horowitz M., Rau G. H., 1996, *Mid-Pliocene warmth: stronger greenhouse stronger conveyor*, Mar. Micropaleontol., 27 (1–4), 313–326, [http://dx.doi.org/10.1016/0377-8398\(95\)00048-8](http://dx.doi.org/10.1016/0377-8398(95)00048-8).
- Raymo M. E., Ruddiman W. F., 1992, *Tectonic forcing of late Cenozoic climate*, Nature, 359, 117–122, <http://dx.doi.org/10.1038/359117a0>.
- Rodgers K. B., Latif M., Lugetke S., 2000, *Sensitivity of equatorial Pacific and Indian Ocean watermasses to the position of the Indonesian throughflow*, Geophys. Res. Lett., 27 (18), 2941–2945.
- Ruddiman W. F., Raymo M. E., Martinson D. G., Clement B. M., Backman J., 1989, *Pleistocene evolution: northern hemisphere ice sheets and North Atlantic Ocean*, Paleooceanography, 4 (4), 353–412, <http://dx.doi.org/10.1029/PA004i004p00353>.
- Schmiedl G., Mackensen A., Müller P. J., 1997, *Recent benthic foraminifera from the eastern South Atlantic Ocean: dependence on food supply and water masses*, Mar. Micropaleontol., 32 (3–4), 249–288, [http://dx.doi.org/10.1016/S0377-8398\(97\)00023-6](http://dx.doi.org/10.1016/S0377-8398(97)00023-6).
- Schönfeld J., 1996, *The 'Stilostomella extinction'. Structure and dynamics of the last turnover in deep-sea benthic foraminiferal assemblages*, [in:] *Microfossils and oceanic environments*, A. Moguevsky & R. C. Whatley (eds.), Univ. Wales-Aberystwyth Press, Aberystwyth, 27–37.
- Siesser W. G., Bralower T. J., Tang C., Galun B., 1992, *Late Miocene-Quaternary calcareous nannofossil biomagnetostratigraphy on the Exmouth Plateau, Northwest Australia*, Proceedings of Ocean Drilling Program, Scientific Results Vol. 122, U. von Rad, B. U. Haq et al. (eds.), Ocean Drilling Program, College Station, TX, 677–681.
- Sinha D. K., Singh A. K., Tiwari M., 2006, *Palaeoceanographic and palaeoclimatic history of ODP site 763A (Exmouth Plateau), southeast Indian Ocean: 2.2 Ma record of planktic foraminifera*, Curr. Sci. India, 90 (10), 1363–1369.
- Smart C. W., Thomas E., Ramsay A. T. S., 2007, *Middle-late Miocene benthic foraminifera in a western equatorial Indian Ocean depth transect*:

- paleoceanographic implications*, *Palaeogeogr. Palaeoclimatol.*, 247 (3–4), 402–420, <http://dx.doi.org/10.1016/j.palaeo.2006.11.003>.
- Smith R. L., 1992, *Coastal upwelling in the modern ocean*, [in:] *Upwelling systems: evolution since the early Miocene*, C. P. Summerhayes, W. L. Prell & K. C. Emeis (eds.), *Geol. Soc. Spec. Publ.*, 64, 9–28.
- Smith W., Grassle J., 1977, *Sampling properties of a family of diversity measures*, *Biometrics*, 33 (2), 282–292, <http://dx.doi.org/10.2307/2529778>.
- Srinivasan M. S., Sinha D. K., 1998, *Early Pliocene closing of the Indonesian seaway: evidence from north-east Indian Ocean and Tropical Pacific deep sea cores*, *J. Asian Earth Sci.*, 16 (1), 29–44, [http://dx.doi.org/10.1016/S0743-9547\(97\)00041-X](http://dx.doi.org/10.1016/S0743-9547(97)00041-X).
- Stehli F. G., Webb S. D. (eds.), 1985, *The great American biotic interchange*, Plenum, New York, 532 pp.
- Tchernia P., 1980, *Descriptive regional oceanography*, Pergamon, Oxford, 252 pp.
- Thomas E., Gooday A. J., 1996, *Cenozoic deep-sea benthic foraminifers: tracers for changes in oceanic productivity?*, *Geology*, 24 (4), 355–358, [http://dx.doi.org/10.1130/0091-7613\(1996\)024<0355:CDSBFT>2.3.CO;2](http://dx.doi.org/10.1130/0091-7613(1996)024<0355:CDSBFT>2.3.CO;2).
- Toggweiler J. R., Samuels B., 1995, *Effect of Drake Passage on the global thermohaline circulation*, *Deep-Sea Res.*, 42 (4), 477–500, [http://dx.doi.org/10.1016/0967-0637\(95\)00012-U](http://dx.doi.org/10.1016/0967-0637(95)00012-U).
- Tomczak M., Godfrey J. S., 2001, *Regional Oceanography: an introduction*, [originally published by Pergamon 1994. Retrieved Jan., 2002, from <http://www.es.flinders.edu.au/~mattom/regoc/pdfversion.html>].
- Venkatarathnam K., Biscaye P. E., 1977, *Distribution and origin of quartz in the sediment of the Indian Ocean*, *J. Sediment. Res.*, 47 (2), 642–649.
- Wells P., Wells G., Cali J., Chivas A., 1994, *Response of deep-sea benthic foraminifera to Late Quaternary climate changes, southeast Indian Ocean, offshore western Australia*, *Mar. Micropaleontol.*, 23 (3), 185–229, [http://dx.doi.org/10.1016/0377-8398\(94\)90013-2](http://dx.doi.org/10.1016/0377-8398(94)90013-2).
- Wyrski K., 1971, *Oceanographic atlas of the International Indian Ocean Expedition*, *Nat. Sci. Found. Publ.*, OCE/NSF 86-00-001 Washington, DC, 531 pp.
- Zachos J. C., Pagani M., Sloan L., Thomas E., Billups K., 2001, *Trends, rhythms, and aberrations in global climate 65 Ma to present*, *Science*, 292 (5517), 686–692, <http://dx.doi.org/10.1126/science.1059412>.