Blake Outer Ridge: Late Neogene variability in paleoceanography and deep-sea biota

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ABSTRACT

Carbon isotope and benthic foraminiferal data from Blake Outer Ridge, a sediment drift in the western North Atlantic (Ocean Drilling Program Sites 994 and 997, water depth ~2800 m), document variability in the relative volume of Southern Component (SCW) and Northern Component Waters (NCW) over the last 7 Ma. SCW was dominant before ~5.0 Ma, at ~3.6–2.4 Ma, and 1.2–0.8 Ma, whereas NCW dominated in the warm early Pliocene (5.0–3.6 Ma), and at 2.4–1.2 Ma. The relative volume of NCW and SCW fluctuated strongly over the last 0.8 Ma, with strong glacial-interglacial variability. The intensity of the Western Boundary Undercurrent was positively correlated to the relative volume of NCW. Values of Total Organic Carbon (TOC) were ~1.5% in sediments older than ~3.8 Ma, and not correlated to high primary productivity indicators, thus may reflect lateral transport of organic matter. TOC values decreased during the intensification of the Northern Hemisphere Glaciation (NHG, 3.8–1.8 Ma). Benthic foraminiferal assemblages underwent major changes when the sites were dominantly under SCW (3.6–2.4 and 1.2–0.8 Ma), coeval with the Last Global Extinction’ of elongate, cylindrical deep-sea benthic foraminifera, which has been linked to cooling, increased ventilation and changes in the efficiency of the biological pump. These benthic foraminiferal turnovers were neither directly associated with changes in dominant bottom water mass nor with changes in productivity, but occurred during global cooling and increased ventilation of deep waters associated with the intensification of the NHG.

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1. Introduction

Blake Outer Ridge (BOR) in the westernmost part of the North Atlantic Ocean (Fig. 1) is a sediment drift, adjacent to two important components of the Atlantic Meridional Overturning Circulation: the warm, saline Gulf Stream and the deep Western Boundary Undercurrent (WBUC). The BOR, built-up of fine grained nannofossil-bearing hemipelagic sediments (Paull et al., 1996), has been argued to have formed through interaction between the upper part of the WBUC and the lower part of the Gulf Stream, where it detaches from the continental slope (e.g., Stahr and Sanford, 1999). BOR sediments largely consist of material transported from the Canadian continental margin by the WBUC (Reynolds et al., 1999; Balsam and Damuth, 2000) (Fig. 1).

Presently, the flanks of the BOR above ~3500 m are covered by the Northern Component Waters (NCW), carried by the WBUC to the South, with a density of ~27.88 kg/m³ and a dissolved oxygen concentration of ~6.3 ml/L (Bower and Hunt, 2000). The NCW consists of several water masses, including the Upper North Atlantic Deep Water (UNADW) with Labrador Sea Waters at depths shallower than ~2500 m, and Lower North Atlantic Deep Water (LNADW, or Norwegian–Greenland Sea Overflow Water), between ~2500 and 4000 m (Stahr and Sanford, 1999; Evans and Hall, 2008). At depths greater than ~4000 m, the BOR is covered by Southern Component Waters (SCW), mainly fed by the Antarctic Bottom Water (AABW). This bottom watermass, however, consists of a varying mixture of NCW (up to 90%) and SCW (Stahr and Sanford, 1999), where the southern component has been recirculated in a cyclonic gyre north of the BOR, and therefore has the same flow direction as the overlying LNADW at the BOR (Weatherly and Kelley, 1985).

The BOR is thus an important region in the North Atlantic Meridional Overturning Circulation (MOC), and vital for the latitudinal exchange of heat, salt and water (Raymo et al., 1990; Evans and Hall, 2008). The BOR underlies the periphery of the subtropical central gyre, with weak upwelling supplying nutrients to the phytoplankton. Over time, the margin of the gyre has migrated repeatedly, so that the
paleoceanographic and paleoenvironmental evolution of the BOR. Benthic foraminifera are an important proxy to reconstruct paleoceanographic changes in the deep-sea, reflecting the availability and quality of particulate organic carbon (food particles, specifically labile as compared to refractory components), the seasonality or lack thereof of the food supply, and bottom/pore water oxygen concentration, although factors such as bottom current intensity may also play a role (Sen Gupta and Machain-Castillo, 1993; Loubeure and Fariduddin, 1999; Gooday, 2003; Fontanier et al., 2005; Jorissen et al., 2007). We selected high sedimentation rate (Paull et al., 1996) ODP Holes 994C and 997A on the Blake Ridge to increase our understanding of late Neogene deep-sea paleoceanographic changes. We generated a 7 myr record of benthic foraminiferal census data from Holes 994C and 997A, stable carbon and oxygen data on tests of Cibicides species and Orbitoides umbonatus, and data on the organic carbon content of the bulk sediment from Hole 994C. We compared our data with published records of local primary productivity based on diatoms [Site 997, (Ikeda et al., 2000)], calcareous nanoplankton [Site 994C, (Okada, 2000)], and oxygen and carbon isotopic records of diagenetic carbonate from Hole 994C (Pierre et al., 2000).

2. Materials and methods

ODP Holes 994C (31° 47.139′ N; 75° 32.753′ W; present day water depth 2798.1 m; penetration 703.5 meters below sea floor or mbsf) and 997A (31° 50.588′ N; 75° 28.118′ W; present day water depth 2770.1 m; penetration 434.3 mbsf) were drilled during ODP Leg 164, and are located 9.6 km apart on the crest of the BOR ([Paull et al., 1996], Fig. 1). The sediment accumulation rate of the hemipelagic ooze was high during the late Miocene (average ~11 cm/kyr at 994C and ~8 cm/kyr at 997A) and Pliocene (~12.5 cm/kyr at 994C ~10 cm/kyr at 997A), but during the Pleistocene dropped to ~5.5 cm/kyr at 994C and ~4.6 cm/kyr at 997A (Fig. 2). Disseminated gas hydrate occurs throughout the sedimentary section between ~450 and ~180 mbsf (~5 to ~2.9 Ma) in both holes (Paull et al., 1996). Free gaseous methane is present below 450 mbsf, but sediments above 180 mbsf (~2.9 Ma) are devoid of gas hydrate. On BOR, cold methane seeps have been found at ~2150 m water depth (Van Dover et al., 2003; Robinson et al., 2004). There is, however, no evidence that methane from the gas hydrates reached the sea floor in cold seeps at the location of Sites 994 and 997, thus benthic foraminiferal assemblages probably were not exposed to methane seeps (Paull et al., 1996). The source organic matter of the clathrate methane may date to the Paleogene, much older than the sediments in which the hydrates reside (Fehn et al., 2000).

2.1. Faunal analysis

We analyzed 440 (Hole 994C) and 240 (Hole 997A) sediment samples of 10 cm³ volume. Samples were processed following Gupta and Thomas (1999). Samples were soaked in water with baking soda for 8–10 h. A few drops of hydrogen peroxide (2%) were added to indurated samples in order to improve disaggregation. Wet samples were washed over a 63 μm size sieve, then dry-sieved over a 125 μm sieve. The ~125 μm size fraction was used for microscopic examination and census counts of benthic foraminifera. Processed samples were split into suitable aliquots to obtain about 250–300 specimens of benthic foraminifera per sample. A total of 220 and 160 species were recorded from Holes 994C and 997A, respectively, among which 137 species are common in both holes. Of these, 48 species contribute significantly to the total population (combined from both holes), occurring in more than 100 samples with at least 8% relative abundance in at least one sample. Eighty three species from Hole 994C and 23 species from Hole 997A occur as rare species only, i.e. present in one to five samples at less than 5% relative abundance. Specimens from both sites are generally well preserved, and not

Fig. 1. Location map of ODP Holes 994C and 997A within the oceanographic setting of the Blake Outer Ridge area. Thick solid and dotted lines indicate deep ocean currents and segmented lines indicate surface ocean currents at Blake Outer Ridge (BOR) area, Northwest Atlantic. Thin lines with arrows represent subtropical gyres. Figure is redrawn from report of Shipboard Scientific Party (1998). NADW, AABW and WBUC represent North Atlantic Deep Water, Antarctic Bottom Water and Western Boundary Undercurrent, respectively.
yellowed or highly polished or badly abraded, thus do not show clear signs of reworking. The average age interval is ~15.4 kyr per sample at Hole 994C and ~22.1 kyr at Hole 997A. Separate age models for both sites (Fig. 2) are based on biostratigraphic data (Okada, 2000), as correlated to the numerical time scale of Berggren et al. (1995). Our time resolution cannot resolve precessional and obliquity-paced variability, and our main purpose was to document the longer-time scale paleoceanographic changes.

2.2. Multivariate analysis

Holes 994C and 997A are located close to each other both geographically and bathymetrically, and common species occur at both holes. Therefore, we combined the faunal data from both holes in the statistical analysis. We selected 48 common species with a relative abundance of 8% or more in at least one sample and present in at least 100 samples for factor and cluster analyses using the SAS/STAT package (Appendix 1).

R-mode Principal Component Analysis (PCA) was performed on the correlation matrix followed by an orthogonal VARIMAX rotation to maximize the variance. Based on the scree (x-y) plot of eigen values versus the number of species (variables) and screening of factor scores we retained 9 factors (Table 1, Fig. 3) that account for 47.9% of the total variance. This low variance may be related to the large number of variables (i.e., species) over the studied interval and the large number of samples (680). We used zero to designate the missing values of each species against each observation number in PCA analysis.

We performed Q-mode cluster analysis using Ward’s Minimum Variance method. Prior to cluster analysis, a PCA was performed on the covariance matrix of the 48 highest ranked species from both holes to standardize the dataset. Based on the plot of semi-partial R-squared values versus the number of clusters, nine clusters were identified (Appendix 2). VARIMAX-rotated factors that show high factor scores with well-established species associations were used to identify biofacies. We identified 9 biofacies, and interpreted their paleoenvironments based on present day ecological preferences of the most abundant species in the biofacies (Tables 1 and 2).

2.3. Total Organic Carbon analysis

Total Organic Carbon (TOC) analysis was performed on 0.5 g of finely powdered, oven-dried sediment, which was dissolved in 50 ml water with 20 drops of 1 N HCl solution. Samples were placed for two hours at room temperature on a magnetic stirrer to digest inorganic carbon. Solutions were analyzed using a TOC Analyzer (TOC-VCPH; Shimadzu Corporation, Japan) in the TOC-GC laboratory, Department of Geology and Geophysics, Indian Institute of Technology, Kharagpur. 2 N HCl and 25% phosphoric acid was further added and purged for 1.5 min for complete digestion of inorganic carbon, and to bring the pH of the solution to 2–3. The machine was standardized using a KH Phthalate synthetic standard. The calibration curve was drawn through the scatter readings of 8 standard solutions (10, 25, 50, 100, 200, 300, 400, and 500 ppm). For the analysis of each sample, 2 to 5 injections were chosen, taking the average of two readings with a standard deviation less than 0.1 and a coefficient of variance less than
Table 1

Benthic foraminiferal biofacies with their factor scores and preferred environments at Hole 994C and 997A.

<table>
<thead>
<tr>
<th>Biofacies</th>
<th>% variance</th>
<th>Factor scores</th>
<th>Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Uhc–Ulh (factor 1 + ve)</td>
<td>6.15542</td>
<td>0.76913</td>
<td>High organic carbon, independent of bottom water oxygenation</td>
</tr>
<tr>
<td>Uvigerina hispida</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Uvigerina hispido-costata</td>
<td></td>
<td>0.72611</td>
<td></td>
</tr>
<tr>
<td>Uvigerina peregrina</td>
<td></td>
<td>0.27672</td>
<td></td>
</tr>
<tr>
<td>Globocassidulina tumida</td>
<td></td>
<td>0.22165</td>
<td></td>
</tr>
<tr>
<td>O-Pb (factor 3 + ve)</td>
<td>5.624418</td>
<td>−0.51477</td>
<td>Cosmopolitan, well oxygenated, relatively low carbon</td>
</tr>
<tr>
<td>Pullenia bulbiloides</td>
<td></td>
<td>−0.21869</td>
<td></td>
</tr>
<tr>
<td>Bp–Bp (factor 5 + ve)</td>
<td>5.36951</td>
<td>0.70542</td>
<td>High organic carbon, possibly low oxygen; S. umbonatus belongs to the ‘extinction group’ taxa; miliolids generally indicate well oxygenated environment, however</td>
</tr>
<tr>
<td>Bolivina pseudopunctata</td>
<td></td>
<td>0.48213</td>
<td></td>
</tr>
<tr>
<td>Bolivina paulea</td>
<td></td>
<td>0.36347</td>
<td></td>
</tr>
<tr>
<td>Peneropoma fusiformis</td>
<td></td>
<td>0.26770</td>
<td></td>
</tr>
<tr>
<td>Cibicides bradyi</td>
<td></td>
<td>0.25376</td>
<td></td>
</tr>
<tr>
<td>Pyrgo lucernula</td>
<td></td>
<td>0.25016</td>
<td></td>
</tr>
<tr>
<td>Stilostomella lepidula</td>
<td></td>
<td>0.24580</td>
<td></td>
</tr>
<tr>
<td>Nonionella auris</td>
<td></td>
<td>0.60907</td>
<td></td>
</tr>
<tr>
<td>Gyroidinoides cibensis</td>
<td>5.31194</td>
<td>0.37708</td>
<td>Intermediate to high organic carbon flux and low oxygen, possibly influence of bottom currents</td>
</tr>
<tr>
<td>Pullenia bulbiloides</td>
<td></td>
<td>0.32291</td>
<td></td>
</tr>
<tr>
<td>Cibicides bradyi</td>
<td></td>
<td>0.26764</td>
<td></td>
</tr>
<tr>
<td>Gyroidinoides nitidula</td>
<td></td>
<td>0.25427</td>
<td></td>
</tr>
<tr>
<td>Sphaeroidina bulbiloides</td>
<td></td>
<td>0.25304</td>
<td></td>
</tr>
<tr>
<td>Quinqueloculina weaveri</td>
<td></td>
<td>0.22933</td>
<td></td>
</tr>
<tr>
<td>Globocassidulina subglobosa</td>
<td></td>
<td>0.22933</td>
<td></td>
</tr>
<tr>
<td>Sc–Pa (factor 7 + ve)</td>
<td>5.11329</td>
<td>0.70536</td>
<td>All taxa in this group are ‘extinction group’, seen as indicative of overall medium-high food supply, probably well oxygenated</td>
</tr>
<tr>
<td>Stilostomella consoberina</td>
<td></td>
<td>0.38673</td>
<td></td>
</tr>
<tr>
<td>Pleurostomella alternans</td>
<td></td>
<td>0.34619</td>
<td></td>
</tr>
<tr>
<td>Nodosaria longisata</td>
<td></td>
<td>0.26418</td>
<td></td>
</tr>
<tr>
<td>Denitella stimulae</td>
<td></td>
<td>0.25034</td>
<td></td>
</tr>
<tr>
<td>Fa–Rg (factor 10 + ve)</td>
<td>5.01030</td>
<td>0.73723</td>
<td>Low to intermediate organic carbon</td>
</tr>
<tr>
<td>Prionodinaria advena</td>
<td></td>
<td>0.49046</td>
<td></td>
</tr>
<tr>
<td>Robulus gibbus</td>
<td></td>
<td>0.26040</td>
<td></td>
</tr>
<tr>
<td>Glandulina laevigata</td>
<td></td>
<td>0.25004</td>
<td></td>
</tr>
<tr>
<td>Si–Mg (factor 4 – ve)</td>
<td>5.45012</td>
<td>−0.52603</td>
<td>Intermediate organic flux, possibly refractory organic carbon (Stilostomella lepidula and Pleurostomella alternans are extinction group of species)</td>
</tr>
<tr>
<td>Stilostomella lepidula</td>
<td></td>
<td>−0.22263</td>
<td></td>
</tr>
<tr>
<td>Melonis pompiroides</td>
<td></td>
<td>−0.20075</td>
<td></td>
</tr>
<tr>
<td>Astrononian umbilicatum</td>
<td></td>
<td>−0.02902</td>
<td></td>
</tr>
<tr>
<td>Pleurostomella alternans</td>
<td></td>
<td>−0.23902</td>
<td></td>
</tr>
<tr>
<td>Gp–Au (factor 12 + ve)</td>
<td>4.90002</td>
<td>0.76180</td>
<td>High organic carbon, potentially low oxygen; refractory organic carbon</td>
</tr>
<tr>
<td>Globobulimina pacifica</td>
<td></td>
<td>0.34879</td>
<td></td>
</tr>
<tr>
<td>Astrononian umbilicatum</td>
<td></td>
<td>0.24438</td>
<td></td>
</tr>
<tr>
<td>Pullenia quinqueloba</td>
<td></td>
<td>0.77873</td>
<td>Intermediate organic carbon flux, possibly with high seasonality</td>
</tr>
<tr>
<td>Cc–Ba (factor 11 + ve)</td>
<td>4.99278</td>
<td>0.53090</td>
<td></td>
</tr>
<tr>
<td>Cassidulina carinata</td>
<td></td>
<td>0.53090</td>
<td></td>
</tr>
<tr>
<td>Bulimina alazenseis</td>
<td></td>
<td>0.53090</td>
<td></td>
</tr>
</tbody>
</table>

2% as the final value. TOC analysis were performed on 447 samples from Hole 994C.

2.4. Stable isotope analysis

Five to ten individuals of Oridorsalis umbonatus, a common species present in most samples, were picked from 191 samples from Hole 994C for stable carbon and oxygen isotopic analysis. In samples where O. umbonatus specimens were rare, Cibicides wuellerstorfi, Cibicides kullenbergi and/or Cibicides bradyi were analyzed. Pre-cleaned samples were reacted in 100% orthophosphoric acid at 70 °C using a Finnigan-MAT Kiel III carbonate preparation device. Evolved CO2 gas was measured online with the Finnigan-MAT 252 Mass Spectrometer at the University of Florida, with standard NBS-19 for calibration. Isotopic results are reported in standard delta notation relative to Vienna Pee Dee Belemnite (VPDB). Analytical precision is estimated (1 standard deviation of standards run with samples) to be ±0.02‰ for δ13C and ±0.07‰ for δ18O (n=58). To correct for the isotopic offset of oxygen and carbon isotope values (vital/habitat effect) between the shallow infaunal O. umbonatus and the epifaunal C. wuellerstorfi, stable isotope values were adjusted to C. wuellerstorfi using the scale δ13C of C. wuellerstorfi = δ13C of O. umbonatus + 1‰ and δ18O of C. wuellerstorfi = δ18O of O. umbonatus − 0.5‰. (Shackleton et al., 1984).

The δ18O values of C. wuellerstorfi were adjusted to equilibrium with sea water by adding 0.64‰. (Shackleton et al., 1984; Zachos et al., 2001).

2.5. SEM study

Diagenetic carbonate nodules are common in both holes (Pierre et al., 2000). The formation of diagenetic calcite may be responsible for changes in oxygen and carbon isotopic values because the foraminiferal tests may contain post-depositional, authigenic carbonates. In gas hydrate containing sediments as recovered at Sites 994 and 997, authigenic calcite may have highly depleted δ18O values [around −20‰, (e.g., Millo et al., 2005)]. To examine whether benthic foraminifera have undergone diagenetic changes, SEM photographs of broken specimens of a few species from Hole 994C were taken from 10 randomly picked samples (Fig. 4, Table 3).

3. Results

3.1. Biofacies distribution

Multivariate analysis of faunal data helps to remove noise from the data set induced by post-mortem taphonomic processes and make it
Fig. 3. Benthic foraminiferal biofacies plotted against age and combined with cumulative percentages of major species of each biofacies at Holes 994C (black) and 997A (red). Full forms of each abbreviated biofacies are given in Table 1.
Table 2
Inferred ecological preferences of different benthic foraminifera used in the present study.

<table>
<thead>
<tr>
<th>Benthic foraminifer species</th>
<th>Ecological preferences</th>
</tr>
</thead>
<tbody>
<tr>
<td>Astronion umbilicatum</td>
<td>Not conclusive: associated with high and sustained organic flux (Hald and Korsun, 1997; Gupta and Thomas, 1999; Gupta et al., 2006a,b), alternatively with low primary productivity, well ventilated water column and high salinity (Singh and Gupta, 2004). Possibly, like Melonis, using refractory organic carbon (Caralp, 1989); other Astronion species are generally thought to be shallow infaunal (Rasmussen et al., 2002).</td>
</tr>
<tr>
<td>Bolivina paula</td>
<td>Species of Bolivinids in general indicate dysoxic, organic-rich environments (Sen Gupta and Machain-Castillo, 1993; Wefers et al., 1994; Hill et al., 2003). Common on continental margins, e.g. under WBLS in southern Atlantic (de Mello e Sousa et al., 2006).</td>
</tr>
<tr>
<td>Bolivina pseudopunctata</td>
<td>Organically enriched environments, possibly oxygen depleted (Sen Gupta and Machain-Castillo, 1993; Rathburn and Corliss, 1994; Gooday, 2003). Used as a proxy for NADW flux and treated as a &quot;warm benthos fauna&quot; restricted to interglacial stages (Schmiedl and Mackensen, 1997).</td>
</tr>
<tr>
<td>Bulimina alazanensis</td>
<td>Opportunistic taxa (Nees and Struck, 1999), C. carinata and Gyroidioides nitidula association indicative of intermediate organic flux and intermediate to high seasonality (Gupta and Thomas, 2003), may survive at low oxygen (Jorissen et al., 2007).</td>
</tr>
<tr>
<td>Cassidulina carinata</td>
<td>Not conclusive: said to be intolerant to low oxygen conditions (Denne and Sen Gupta, 1991; Barmawi-jaja et al., 1992), and associated with well-oxygenated deep water with low organic flux and high seasonality (Singh and Gupta, 2004). In contrast, pore patterns and rounded periphery of the test said to indicate adaptation to lower oxygen levels (Rathburn and Corliss, 1994). Co-occurrence with Bolivina species suggests tolerance to lower oxygen levels and also occurs with bolivinids on southern Atlantic continental margin (de Mello e Sousa et al., 2006).</td>
</tr>
<tr>
<td>Cibicides bradyi</td>
<td>Opportunistic (Alve, 1995; Rasmussen et al., 2002), tolerant to low oxygen and survive in anoxia (Alve, 1995; Bernard and Alve, 1996; Gustafsson and Nordberg, 2001); abundant in cold-seep environments (Rathburn et al., 2000).</td>
</tr>
<tr>
<td>Dentalina stimulae</td>
<td>Deep-infaunal, low oxygen, high organic carbon environment (Sen Gupta and Machain-Castillo, 1993; Gooday, 2003); more abundant in sediments with refractory, more degraded organic matter (Schmiedl et al., 2000); may be indicative of laterally transported, partially degraded organic matter (Fontanier et al., 2005).</td>
</tr>
<tr>
<td>Globocassidulina subglobosa</td>
<td>Cosmopolitan species, oligotrophic (Singh and Gupta, 2004 and references therein); abundance related to increased vigor of bottom currents (Schmiedl et al., 1997; Rasmussen et al., 2002; Smart et al., 2007).</td>
</tr>
<tr>
<td>Globocassidulina tumida</td>
<td>Not well known. Globocassidulina species are common in regions where organic carbon flux is very high throughout the year (Singh and Gupta, 2004), and have higher abundance in the area of increased bottom water currents (Schmiedl et al., 1997; Rasmussen et al., 2002; Smart et al., 2007).</td>
</tr>
<tr>
<td>Gyroidioides cibaoensis</td>
<td>Low oxygen (Gupta and Thomas, 1999; Gupta et al., 2008), food limited or pulsed food supply (Mackensen et al., 1995; De Rijk et al., 1999), oligotrophic (Singh and Gupta, 2004).</td>
</tr>
<tr>
<td>Gyroidioides nitidula</td>
<td>Resembling G. arculis, found in a food limited environment (Singh and Gupta, 2004 and references therein).</td>
</tr>
<tr>
<td>Melonis pumillosa</td>
<td>Moderate productivity and intermediate seasonality (Gupta and Thomas and references therein), refractory organic carbon (Caralp, 1989; Fontanier et al., 2005).</td>
</tr>
<tr>
<td>Nonionella auris</td>
<td>Survives low oxygen, even anoxic conditions, occurs H2S-containing sediments, cold-seep environments, feed on bacteria (Wefer et al., 1994); generally abundant under high productivity (Gooday, 2003). Some Nonionella species (N. stella) have been reported to grow fast under very high productivity (Corliss and Silva, 1993), even at low oxygen conditions (Bernard et al., 1997).</td>
</tr>
<tr>
<td>Odorarlis umbonatus</td>
<td>Cosmopolitan, very long-lived taxa (Gupta and Thomas, 1999). Well oxygenated, low organic carbon environment (Mackensen et al., 1985; Gooday, 2003), organic food limited and low oxygen (Rathburn and Corliss, 1994). Probably environmentally flexible, occurs over wide depth range and age range (since Late Cretaceous), (Kaiho, 1998). Association with P. bulloides indicates they can survive with low oxygen and intermediate to high organic carbon rich environment.</td>
</tr>
<tr>
<td>Pullenia quinquevalva</td>
<td>High organic matter (Schnitker, 1986) and deposit feeder (Liu et al., 1997).</td>
</tr>
<tr>
<td>Pyrgo lucernula and</td>
<td>Member of miliolids may prefer cool, oligotrophic, well-oxygenated bottom water conditions (Mackensen et al., 1995; Altenbach et al., 1999; Jorissen et al., 1999; Schmiedl et al., 2000; Rasmussen et al., 2002; Gooday, 2003; Gupta and Thomas, 2003). Quinqueloculina and Pyrgo are also reported in seep environments (Robinson et al., 2004). Some species of miliolids (e.g. Articulina spp.) are associated with more oxygenated conditions during recovery from low oxygen (e.g. Mediterranean sapropels, (Mullineaux and Lohmann, 1981; Jorissen, 1999)).</td>
</tr>
<tr>
<td>Quinqueloculina weaveri</td>
<td>Not well constrained. Infaunal (Corliss, 1991), another species of this genus, R. iota is a characteristic of oxygen minimum zones (Hermelin and Shimmield, 1990).</td>
</tr>
<tr>
<td>Uvigerina hispida</td>
<td>In general uvigerinds show close affinity with high productivity independent of bottom water oxygenation (Lutze, 1986; Rathburn and Corliss, 1994). U. hispida shows higher abundance in low oxygen, high organic carbon rich environment, indicates period of erosion and downward transportation (McDougall, 1996).</td>
</tr>
<tr>
<td>Uvigerina hispido-costata</td>
<td>U. hispido-costata is abundant in high organic carbon flux, low oxygen settings (Gupta and Thomas, 2003; Murgese and Deckker, 2005).</td>
</tr>
<tr>
<td>Uvigerina peregrina</td>
<td>Uvigerina peregrina is more closely related to the continuous organic carbon flux than to the oxygen minima (Rathburn and Corliss, 1994), association of U. peregrina with B. aculeata, C. bradyi, U. hispida and U. proboscidea indicates presence of NADW (Murray, 2006).</td>
</tr>
<tr>
<td>Uvigerina subglobosa</td>
<td>This group belongs to the elongated benthic community and disappears during the last global extinction (mid Pleistocene transition) except Glandulina laevigata. Moderately deep infaunal, tolerant of low oxygen (Gupta, 1993) and wide range of bottom water temperature and dissolve oxygen (Hayward et al., 2007), found in oligotrophic and eutrophic region with sustained or highly seasonal phytoplankton productivity, tolerant of changes in the quantity or seasonality of organic carbon (Hayward et al., 2010a).</td>
</tr>
</tbody>
</table>

possible to identify significant associations of species. We identified nine biofacies (Table 1), indicated by the abbreviated names of their dominant species. We recognized: 1. Uhc-Uh (dominant species: Uvigerina hispido-costata, Uvigerina hispida, Uvigerina peregrina and Globocassidulina tumida), 2. Sc-Pa (dominant species: Stilostomella consorbnia, Pleurostomella alternans, Nodosaria longisca and Dentalina stimulae), 3. GC-Pb (dominant species: Gyroidioides cibaoensis, Pullenia bulboides, Cibicides bradyi, Gyroidioides nitidula, Sphaeroidina bulboides, Quinqueloculina weaveri and Globocassidulina subglobosa), 4. Fa-Rg (dominant species: Frondicularia advena, Robulus gibbus and Glandulina laevigata), 5. Ou-Pb (dominant species: Oridorsalis umbonatus and P. bulboides), 6. Bp-Bp (dominant species: Bolivina pseudopunctata, Bolivina paula, Fursenkoina fusiformis, C. bradyi, Pyrgo lucernula, Stilostomella lepidula and Nonionella auris), 7. Sl-Mp (dominant species: S. lepidula, Melonis pumillosa, Astrongion umbilicatum and Pleurostomella alternans), 8. Cc-Ba (dominant species: Cassidulina carinata and Bulimina alazanensis) and 9. Gp-Au (dominant species: Globobulimina pacifica, Astrongion umbilicatum and Pullenia quinqueloba). All species in biofacies 2 (Sc-Pa) are elongate cylindrical species, as is S. lepidula (Fig. 3).
Over the full studied interval, the benthic foraminiferal assemblages are typical of continental margin regions, with common species indicative of a fairly high food supply, possibly mixtures of more labile, locally produced organic matter and laterally transported, more refractory organic material (e.g., Fontanier et al., 2005; de Mello e Sousa et al., 2006; Jorissen et al., 2007; Thomas, 2007). The sites were probably influenced by a fairly high but fluctuating food supply, with the Ou–Pb biofacies indicative of the somewhat lower productivity periods, the Uhc–Uh and Bp–Bp biofacies indicative of more productive intervals.

Biofacies Ou–Pb and Uhc–Uh are present throughout the studied interval, with the highest abundances of Ou–Pb before 5 Ma, and those of Uhc–Uh between 5.3 and 2 Ma. Biofacies Bp–Bp was present throughout the studied interval, but at significant values only after 6.2 Ma, and it increased in abundance after 5 Ma and again after 1.8 Ma. Biofacies Fa–Rg was overall rare, but least so between 5.2 and 2.6 Ma. Biofacies Gp–Au and Sl–Mp became common at 3.8–3.7 Ma, Cc–Ba at 2.7 Ma. Biofacies Sc–Pa and Gc–Pb (in Hole 994C) became much less abundant at 2.5 and 3.1 Ma, respectively, Gp–Au and Sl–Mp at 0.8 and 1.0 Ma, respectively (Fig. 3).

There was a major faunal turnover in the interval between 3.8 Ma and 2.5 Ma: Biofacies Sl–Mp, Gp–Au, and Cc–Ba replaced biofacies Gc–Pb, Sc–Pa and Fa–Rg. A smaller turnover occurred at 1.0–0.8 Ma. The data on overall turnover in benthic assemblages from this sediment drift on a continental margin resemble data from open ocean sites and other oceans, in that we see a disappearance of elongate species with a complex aperture at times of increased intensity of glaciation (e.g., Kawagata et al., 2005; Hayward et al., 2010a,b). There is no overall clear change in assemblages indicative of a major increase or decrease in food supply over the studied interval, although the type of organic matter may very well have changed over time, as indicated by the turnover in biofacies (Tables 1, 2; see below).

3.2. Geochemical data

3.2.1. Diagenetic effects

The samples studied were deposited over a large burial range, with the lowermost samples in Hole 994C now at about 700 mbsf (Fig. 2).
There is considerable evidence for recrystallization of material in the holes, and the presence of gas hydrates shows that diagenesis of the sediments has been fairly intensive, at least in some depth intervals. The first question we need to answer thus is whether the benthic foraminiferal stable isotope values reflect the environment of deposition of the sediments or the downhole diagenesis.

We examined specimens of benthic foraminifera using Scanning Electron Microscopy, breaking specimens to evaluate whether diagenetic calcite was commonly present. Although most specimens show at least some recrystallization, we found no evidence for severe overgrowth in most specimens investigated. The exceptions are a few specimens of Cibicides bradyi (3.52 Ma) and Oridorsalis umbonatus (5.08 Ma), showing calcite overgrowths inside their tests (Fig. 4). The overgrowths appear to be cryptocrystalline and patchily distributed, in contrast with other locations.

We thus argue that our records overall reflect paleoceanographic conditions rather than diagenetic processes, with the more negative values of the δ13C values of Cibicides wuellerstorfi from these two samples are 0.76‰ and −1.53‰, i.e., they are isotopically heavier than samples without overgrowth (Table 3). The δ13C and δ18O values of diagenetic calcite from Hole 994C (Pierre et al., 2000) do not show a significant correlation with the carbon and oxygen isotopic values of benthic foraminifera from the same levels (Fig. 5). The plot of δ18O vs. TOC values shows a significant negative correlation (R = −0.5), and there is no significant correlation between δ13C and TOC values.

We compared our benthic foraminiferal isotope records with the data in the global compilation of Zachos et al. (2001) (Fig. 6a) and Cramer et al. (2009) (Fig. 6b), as well as with those from a short time interval at close-by Site 1058 (Franz and Tiedemann, 2002). By far most values in our δ18O records are within the variability of the data presented by these two groups of authors, with only a few exceptions. We excluded some extreme data points that differ by more than 0.5‰ from the compilation (indicated by bold, Appendix 3). The δ13C values for most samples in our records overlap the range of values in Zachos et al. (2001), being between −1.5‰ and +1‰. We excluded the few values outside this range from further analysis (bold face, Appendix 3). For some time intervals (see below), our values are significantly lower (1 to 3%) than the coeval values in the records in Zachos et al. (2001), although still within the −1.5‰ to +1‰ range. These lower values overlap with values for SCW as shown in Franz and Tiedemann (2002) and Poore et al. (2006). Our data are less similar to the values published by Cramer et al. (2009) for various ocean basins, but are closest to these authors’ data from the Southern North Atlantic (Fig. 6b). Neither Zachos et al. (2001) nor Cramer et al. (2009) include data from BOR. The lower δ13C values in our records, which overlap with values for Southern Component Waters in Franz and Tiedemann (2002) and Poore et al. (2006) occur at earlier times (i.e., back to the beginning of our records at 7 Ma) than in the compiled records of Zachos et al. (2001) and Cramer et al. (2009) (Fig. 6a,b), indicating that such SCW occurred in the westernmost North Atlantic Basin over the whole period studied, in contrast with other locations.

We thus argue that our records overall reflect paleoceanographic conditions rather than diagenetic processes, with the more negative values of the δ13C record reflecting mainly SCW, the more positive ones NCW, after excluding some extreme values, which we consider affected by diagenetic processes (Appendix 3; bold face). Our time resolution makes it impossible to draw inferences about changes on glacial–interglacial time scales, and we chose to take 5-point moving averages of the isotope data and TOC to make the overall trends clear (Fig. 7).

### 3.2.2. TOC and stable isotope data

Total Organic Carbon values at Hole 994C fluctuate, but are generally highest in the lower part of the record (before ~3.6 Ma), with an average of 1.5 (wt%). The values declined until about 1.6 Ma, after which they remained stable at around 0.75 wt%. Comparison of Fig. 3 with Figs. 7 and 8 shows that biofacies Uhc–Uc, Sc–Po, Cc–Pb, Fa–Pg and Ouc–Pb dominated intervals of high TOC, whereas biofacies Bp–Bp, Sl–M, Cc–Ba and Gp–Au were dominant during intervals of lower TOC values.

The pattern of TOC is similar (though opposite in sign) to the record of oxygen isotopic values of benthic foraminifera: these values...
fluctuated around 3.25‰ before about 3.6 Ma, then increased until about 1.6 Ma, and afterwards fluctuated around a value of about 4‰. There thus is a clear, negative correlation between these two parameters (Fig. 5). Our records are not of sufficient time resolution to document the details of the glacial–interglacial variability over the time period investigated, but document the overall well-known record of intensification of NHG, and show that this increased glaciation was linked to decreasing deposition of TOC at BOR.

The benthic carbon isotope values are not significantly correlated to TOC and δ18O values (Fig. 5) of diagenetic carbonate. The record shows great variability, but overall high values (N −0.5‰) between 5.0 and 3.6 Ma, between 2.4 and 1.2 Ma, and between 0.8 Ma and the last few ten thousands of years (Fig. 7). The relative abundance of Cibicides wuellerstorfi shows overall higher values (>10%) in intervals with higher δ13C values (Figs. 7, 8).

4. Discussion

The climate of the Earth has cooled over the last 7 Ma (e.g., Zachos et al., 2001), with significant build up of ice on Southern Greenland in the late Miocene (Maslin et al., 1998; Haug et al., 2005), but global cooling was interrupted by warm phases such as the early Pliocene (e.g., Wara et al., 2005; Pagani et al., 2010), during which transient changes occurred in the size in Antarctic ice sheets (e.g., Pollard and deConto, 2009). Overall, however, there has been a net increase in polar ice volume since the late Miocene, and an increase in the magnitude of glacial–interglacial variability, with the 100 kyr eccentricity-forced component of orbitally-driven glacial–interglacial climate variability dominant over the last 800–900 kyr (e.g., Gupta et al., 2001; Haug et al., 2005).

In the North Atlantic, the production of the NCW was strongly enhanced during the Pleistocene warm interglacial intervals (e.g., Reynolds et al., 1999; Frank et al., 2002; Franz and Tiedemann, 2002; Hagen and Keigwin, 2002; Lynch-Stieglitz et al., 2007; Evans and Hall, 2008). During glacial intervals, in contrast, the upper boundary between SCW and the glacial equivalent of NADW [GNAIW; (Lynch-Stieglitz et al., 2007)] shoaled by more than 2200 m along BOR (Franz and Tiedemann, 2002; Evans and Hall, 2008). Less is known of the variability in NCW/SCW for earlier periods.

The δ13C values of Cibicides wuellerstorfi have been widely used to detect changes in deep-water ventilation in the Atlantic and between oceans (e.g., Haug and Tiedemann, 1998). In general, δ13C values of benthic foraminiferal tests (commonly Cibicides spp.) which calcified within the poorly-ventilated, nutrient-rich SCW are more depleted in δ13C DIC, with values between 0 and −1‰. Tests calcified within NCW are relatively enriched, with values >0‰ (Kroopnick, 1985; Raymo et al., 1998; Franz and Tiedemann, 2002; Curry and Oppo, 2005; Lynch-Stieglitz et al., 2007; Ravelo and Hillaire-Marcel, 2007). Very large fluctuations in δ13C values thus are seen at locations where SCW alternated with NCW (e.g., Franz and Tiedemann, 2002).

Due to the time resolution of our study we cannot make inferences about changes on glacial–interglacial time scales, but the carbon isotope and Cibicides wuellerstorfi % data (Fig. 8) indicate that our sites were dominantly covered by SCW (characterized by low δ13C values...
and low % of C. wuellerstorfi) between 7 and 5 Ma (late Miocene–earliest Pliocene), with $\delta^{13}C$ values below those in the Zachos et al. (2001) curves (Fig. 6), and overlapping with SCW values in Poore et al. (2006). Calcareous nannoplankton and diatom assemblages (Ikeda et al., 2000; Okada, 2000) indicate fairly high productivity at that time, in agreement with dominance by benthic foraminiferal biofacies (biofacies Gc–Pb, Sc–Pa and Uhc–Uh), indicative of a high and probably not highly seasonal food supply, with possible current influence. The overlying surface waters were probably mainly gyre-margin environments, with some upwelling of the nutrient-enriched SCW, which reached a few km depth (Ikeda et al., 2000; Okada, 2000). This interval in the late Miocene and earliest Pliocene was characterized by the presence of a western as well as eastern Antarctic ice sheet (e.g. Zachos et al., 2001; Cramer et al., 2009), so that the volume of SCW may have been high due to cooling at high southern latitudes, whereas NCW volume may have been limited by the relatively shallow depth of the sill in the northernmost Atlantic over which these NCW waters flow in the Atlantic Ocean (Wright and Miller, 1996; Poore et al., 2006).

Between about 5.0 and 3.6 Ma the sites were mainly under the influence of NCW, with benthic $\delta^{13}C$ values of up to 1.25‰. During this period the diatom and nannofossil productivity declined (Fig. 8), although the TOC in the sediments shows a moderate change only. The benthic foraminiferal biofacies underwent minor changes only, with the more common presence of biofacies Fa–Rg, possibly indicative of a somewhat lower food supply, or possibly somewhat less labile, more refractory organic matter, possibly also leading to the somewhat increased abundance of biofacies Bp–Bp. This organic matter might have arrived by lateral transport of the vigorous WBUC, with higher current intensity indicated by the higher % of Cibicides wuellerstorfi (Figs. 7, 8). Possibly, the younger NCW brought fewer nutrients to BOR, resulting in lesser primary productivity at the surface, or the gyre location shifted, bringing more oligotrophic waters over the BOR during the warm early Pliocene (Wara et al., 2005; Dowsett et al., 2009; Seki et al., 2010). This overall warm period thus may have seen Atlantic MOC similar to that of the present day, with large volume NCW production, relatively low primary productivity in water overlying BOR. NCW may have formed at a similar

![Fig. 6. a: Benthic oxygen and carbon isotope values of Hole 994C compared to the global compilation of Zachos et al. (2001). Gray bars indicate inferred times of SCW and NCW dominance at the sites. b: Benthic oxygen and carbon isotope values of Hole 994C compared to the compilation of Cramer et al. (2009) from South Atlantic and Northern part of the Southern Ocean. Gray bars indicate inferred times of SCW and NCW dominance at the sites.](image-url)
location as today’s NADW, i.e. relatively further North than during colder intervals, and in addition the elevation of the sill where these waters flow into the Atlantic Ocean to form the WBUC may have been lowered (Wright and Miller, 1996; Poore et al., 2006).

The period between 3.6 and 2.4 Ma saw a return of SCW to the sites, with benthic δ¹³C values between −1 and −1.5‰. This was the time of major intensification of the Northern Hemisphere Glaciation (Haug and Tiedemann, 1998; Zachos et al., 2001; Mudelsee and Raymo, 2005), as indicated at our sites by increasing benthic δ¹⁸O values, as well as declining TOC values. The decline in TOC values was likely not simply related to the change in dominant deep-water mass over the sites, since the change from SCW to NCW at about 5 Ma did not have a significant effect on TOC. This interval also saw a decline in depth of the Iceland–Greenland sill (Poore et al., 2006), possibly limiting the volume on NCW. This interval also saw the largest turnover of benthic biofacies, with the decrease in Fa–Rg (low to intermediate food supply), Sc–Pa (low to intermediate food, extinction group species), and Gc–Pb (intermediate-high food flux), and the increase in biofacies Bp–Bp (high organic carbon), SI–Mp (intermediate-refractory food), Gp–Au (high food supply, possibly refractory), and Cc–Ba (intermediate food flux, possibly seasonal). There thus may have been an overall further increase in the flow of more refractory organic carbon to the sites (dominance of biofacies SI–Mp), combined with a more seasonal food flux. The primary productivity of diatoms and calcareous nannoplankton increased once more, similar to its status at 7.0–5.0 Ma, whereas TOC started to decline, in contrast to that earlier period. Possibly, less of the more labile organic matter reached the sea floor, even with the presence of the less-ventilated SCW, with a change to more refractory and more seasonal food supply. It is also possible that the cooling resulted in increasingly vigorous bottom currents (continuing into the following time periods), leading to less deposition of fine-grained organic material from higher northern latitudes, although this is not confirmed by the overall low abundance of Cibicides wuellerstorfi.

The biofacies which decreased at this time include typical ‘extinction group’ species (Kawagata et al., 2005; Hayward et al., 2010a,b), which became extinct globally during the late Pleistocene cooling of the deep-sea. These species all have complexly structured apertures, suggesting that they may have shared a mode of feeding which no longer exists in the cold, well-ventilated oceans of the Present (Hayward et al., 2010a,b). They may have fed on a phytoplankton source which became extinct during the taxonomic
Waters over the sites were dominated by NCW between 2.4 and 1.2 Ma, a period which saw fairly stable benthic foraminiferal assemblages, despite the change in dominant water mass. In contrast to the period of NCW between 5.0 and 3.6 Ma, this period saw persistent high productivity by diatoms and nannofossils, but TOC remained low. Possibly, the gyre margin remained over the sites. The surface water circulation in the western North Atlantic may have differed from that during the earlier NCW-period, because of the shoaling of the Panamanian isthmus at around 4.6 Ma (Haug and Tiedemann, 1998), strengthening the Gulf Stream, thus keeping the gyre margin further out.

Between 1.2 and 0.8 Ma the SCW returned over the sites, with persistent high productivity of calcareous nannofossils and diatoms, possibly explaining the high abundance of biofacies Bp–Bp. Towards the end of this period climate variability increased, with the establishment of the dominant 100 kyr variability (e.g., Maslin et al., 1998; McClymont and Rosell-Melé, 2005). Benthic assemblages underwent additional turnover during this period, with a decrease in biofacies Sl–Mp (intermediate-refractory food), Gp–Au (high food

Fig. 7. Values of TOC, δ¹³C and δ¹⁸O of C. wuellerstorfi at Hole 994C and relative abundance of C. wuellerstorfi, with interpretation of the presence of Southern and Northern Component Waters over the sites. Thick lines represent 5 point moving average.
supply, possibly refractory). The first of these reflects the regional expression of the global extinction of the cylindrical species, the last of which became extinct at this time (Hayward et al., 2010a,b). The second decline suggests an overall decline in transport of more refractory organic matter to the sites. Such a decline might be linked to increasingly vigorous bottom currents (as argued above), or declining refractory organic matter transported from land due to increasing glaciation and declining vegetation on land in the north (with both NCW and SCW flowing over BOR from the North).

The last 800 kyr is characterized by strongly variable conditions, with declining diatom and nannofossil productivity, and strong fluctuations in NCW–SCW over the sites, due to the high intensity of climate variability at the 100 kyr periodicity, and benthic assemblages were similar to those in present days.
5. Conclusions

Environments of the Blake Outer Ridge at depths of ~2800 m have been alternatively influenced by dominant Northern Component Waters and Southern Component Water over the last 7 Ma. This alternating influence has not been shown clearly in records from other locations in the North Atlantic, probably because these are further away from the Western Boundary Undercurrent (WBUC), which transport most of the NCW to the South.

The late Miocene through the earliest Pliocene (7–5 Ma) saw the sites mostly covered by SCW, at overall high productivity under a gyre margin. During the warmer early Pliocene (5.0–3.6 Ma) NCW water was present at the sites, SCW was present between 3.6 and 2.4 Ma during the increase of the Northern Hemisphere Glaciation, NCW returned between 2.4 and 1.2 Ma, and SCW between 1.2 and 0.8 Ma, followed by strongly fluctuating conditions. Dominance of NCW or SCW is linked to high latitude climate, with more NCW forming during overall warmer periods, as well as to the elevation of the Greenland–Iceland sill, with lesser elevation leading to larger volumes of NCW.

There is no clear correlation between high primary productivity and presence of NCW/SCW at the site locations, and there is no clear correlation between TOC and primary productivity indicators. Possibly some of the organic matter preserved was refractory organic matter from lateral transport or transport from the continental margin.

With increasing Northern Hemisphere Glaciation (3.6–2.4 Ma) the TOC of the sediments declined at the same time as the decline in temperature/increase in ice volume (increasing δ18O values), even as primary productivity (diatoms, nannoplankton) remained high. Possibly more vigorous currents resulted in declining deposition of refractory, fine-grained organic matter to the seafloor, indicating some decoupling between benthic-pelagic processes.

Benthic biofacies do not show strong changes during times of change in bottom water masses over the sites, and also do not show major changes linked to local/regional primary productivity variability (variability in gyre location, upwelling intensity, nutrient content of upwelling waters).

Benthic foraminiferal assemblages are mainly influenced by globally recognized events, i.e., the last global extinction of benthic foraminifera during the intensification of the Northern Hemisphere Glaciation and the change to a world dominated by high amplitude 100 kyr climate variability. The exact causes of the faunal changes are not clear: they have been linked to stepped cooling, changing circulation patterns, increased ventilation and changes in oceanic primary productivity and the efficiency of the biological pump.

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Appendix 1. List of high-ranked benthic foraminifera from Holes 994C and 997A used in R-mode Principal Component Analysis and Q-mode cluster analysis

1. Astrononion umbilicatum (Uchio) = Astrononion umbilicatum Uchio, 1952, p. 36, text fig. 1 – Gupta, 1994, pl. 5, fig. 17.

2. Awehea tosta (Schwager) = Nodosaria tosta Schwager, 1866, p. 219, pl. 5, fig. 42 – Hayward, 2002, pl. 1, figs. 7, 8.

3. Bolivina paula (Cushman and Cahill) = Bolivina paula Cushman and Cahill, 1932, Marszalek et al., 1969, fig. 10.

4. Bolivina pseudopunctata (Höglund) = Bolivina pseudopunctata Höglund, 1947, p. 273, pl. 24, fig. 5, pl. 31, figs. 23, 24 – Sarkar et al., 2009, pl. 2, fig. 2.

5. Bulimina alazanensis (Cushman) = Bulimina alazanensis Cushman, 1927, p. 161, pl. 25, fig. 4 – Gupta, 1994, pl. 3, fig. 7.

6. Bulimina costata (d’Orbigny) = Bulimina costata d’Orbigny, 1852, p. 115 – Sarkar et al., 2009, pl. 2, fig. 9.

7. Cassidulina carinata (Silvestri) = Cassidulina laevigata d’Orbigny var. carinata Silvestri, 1896, p. 104, pl. 2, fig. 10 – Gupta, 1994, pl. 2, fig. 10.

8. Cibicides bradyi (Trauth) = Cibicides bradyi Trauth, 1918, p. 665, pl. 95, fig. 5 – Gupta, 1994, pl. 5, figs. 3, 4.

9. Cibicides kullenbergii (Parker) = Cibicides kullenbergi Parker, 1953, p. 49, pl. 11, figs. 7, 8 – Gupta, 1994, pl. 5, fig. 5.


11. Dentalina stimulea (Schwager) = Nodosaria stimulea Schwager, 1866, p. 226, pl. 6, fig. 57 – Hayward, 2002, pl. 2, figs. 34–35.

12. Eggerella bradyi (Cushman) = Verneuilina bradyi Cushman, 1911, p. 54, pl. 2, text figs. 87a–b – Gupta, 1994, pl. 1, fig. 2.

13. Epistominella exigua (Brady) = Pulvinulina exigua Brady, 1884, p. 696, pl. 103, figs. 13, 14 – Gupta, 1994, pl. 4, figs. 18, 19.

14. Frondicularia advena (Cushman) = Frondicularia advena Cushman, 1923, p. 141, pl. 20, figs. 1, 2 – Barker, 1960, pl. 66, figs. 8–12.

15. Fursenkoina fusiformis (Williamson) = Stainforthia fusiformis (Williamson) = Bulimina piooids d’Orbigny Var. fusiformis Williamson 1858, P. 63, pl. 5, figs. 129, 130 – Hermelin and Scott, 1985, pl. 4, fig. 14.

16. Caudulina laevigata (d’Orbigny) = Nodosaria (Caudulina) laevigata d’Orbigny, 1826, p. 252, pl. 1, figs. 1–4 – Sarkar et al., 2009, pl. 4, fig. 20.

17. Globobulimina pacifica (Cushman) = Globobulimina pacifica Cushman, 1927, p. 67, pl. 14, fig. 12 – Gupta, 1994, pl. 3, fig. 10.

18. Globocassidulina obtusa (Williamson) = Cassidulina obtusa Williamson, 1858, p. 69, pl. 6, figs. 143, 144 – Murray, 2006, (as Cassidulina), Fig. 53, No. 12.


20. Globocassidulina tumida (Heron-Allen and Earland) = Cassidulina laevigata d’Orbigny Var. tumida Heron-Allen and Earland 1922, p. 137, pl. 5, figs. 8–10 – Gupta, 1994, pl. 3, figs. 1, 2.

21. Gyroidinoides cibaoensis (Bermúdez) = Gyroidinoides cibaoensis Bermúdez, 1949, p. 252, pl. 17, figs. 61–63 – Sarkar et al., 2009, pl. 5, fig. 6.

22. Gyroidinoides nitidula (Schwager) = Rotalia nitidula Schwager, 1866, p. 263, pl. 7, fig. 110 – Gupta, 1994, pl. 6, fig. 15.

23. Hoeglundina elegans (d’Orbigny) = Rotalia elegans d’Orbigny, 1826, p. 276, no. 54 – Gupta, 1994, pl. 2, figs. 7, 8.

24. Melonis barleeanum (Williamson) = Nonionina barleeanum Williamson, 1858, p. 32, pl. 3, figs. 68, 69 – Gupta, 1994, pl. 6, fig. 1.

25. Melonis pompilooides (Fichtel and Moll) = Nautilus pompilooides Fichtel and Moll 1798, p. 31, pl. 2, figs. a–c – Gupta, 1994, pl. 6, figs. 2, 3.

26. Nodosaria longiscata (d’Orbigny) = Nodosaria longiscata d’Orbigny, 1846, p. 32, pl. 1, figs. 10–12 – Hayward, 2002, pl. 2, fig. 43.

27. Nonionella auris (d’Orbigny) = Valvulina auris d’Orbigny, 1839, p. 47, lám. 2, figs. 15–17 – Hayward et al., 2002, pl. 1, figs. 36–38.

28. Nuttalides unbomifer (Cushman) = Pulvinulinelllumbonifera Cushman, 1933, p. 90, pl. 9, fig. 9 – Gupta, 1994, pl. 5, figs. 14–16.


Kroopnick, P.M., 1985. The distribution of 13C0 f...


