Paleoenvironmental reconstruction of Challenger Mound initiation in the Porcupine Seabight, NE Atlantic

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The understanding of the paleoenvironment during initiation and early development of deep cold-water coral carbonate mounds in the NE Atlantic is currently a focus of international research. The Integrated Ocean Drilling Program (IODP) Expedition 307 drilled the 155 m high Challenger Mound in the Porcupine Seabight (SW off Ireland) in order to investigate for the first time sediments from the base of a giant carbonate mound. In this study we focus in high resolution on 12 m of sediments from Site 1317 encompassing the mound base. The mound initiation and start-up phase coincide with the intensification of the Northern Hemisphere Glaciation (INHG) at around 2.7 Ma. Further carbonate mound development seems to be strongly dependent on rapid changes in paleoceanographic and climatic conditions at the Pliocene–Pleistocene boundary, especially characterized and caused by the interaction of intermediate water masses, the Mediterranean Outflow Water (MOW), the Eastern North Atlantic Water (ENAW) and the influence of Southern Component Water (SCW).

This study is based on well-established proxies such as $\delta^{18}O$ and $\delta^{13}C$ of planktonic (*Globigerina bulloides*) and benthic foraminifera (*Fontbotia wuellerstorfi*, *Discanomalina coronata*, *Lobatula lobatula*, *Lobatula antarctica*, and *Planulina ariminensis*) as well as grain size parameters to identify the paleoenvironmental and paleoecological setting favourable for the initial coral colonization on the mound. Stable oxygen and carbon isotope records of benthic foraminiferal species indicate that *L. lobatula* provides a reliable isotopic signature for paleoenvironmental reconstructions. In particular, $\delta^{18}O$ values of *L. lobatula* indicate that initial mound growth started in a glacial mode with moderate excursions in $\delta^{18}O$ values. Carbon isotope values of *D. coronata* are significantly offset compared to other epibenthic species. This offset may be related to vital effects. Bottom water temperatures, calculated using standard equations based on $\delta^{18}O$ of foraminiferal tests, range between 7 and 11 °C, consistent with the known temperature range conducive for cold-water coral growth and development.

Bottom currents transporting intermediate water masses of southern origin (Mediterranean and Bay of Biscay) enhanced at 2.6 Ma supporting first coral settlements with the INHG. The benthic $\delta^{13}C$ and the sortable silt records indicate that the early Pleistocene hydrodynamic regime was characterized by weaker current intensities associated with vertical movements of MOW or its replacement by SCW at intermediate depth. After these sluggish phases enhanced MOW flow dominated again and led to stronger current intensities and most probably sediment erosion on Challenger Mound. Erosion in combination with early diagenetic (oxidation) processes overprinted the sediment layers as indicated by dissolved coral skeletons, the increase in Ca-content and sediment density, minimum $\delta^{13}C_{planktonic}$ values, as well as the occurrence of gypsum and pyrite, implying a careful evaluation of original and overprinted geochemical signals. We conclude that the Challenger Mound development was already influenced by short-term variability of water masses from southern origin and possible erosional events comparable to the late Pleistocene setting.

1. Introduction

The European continental margin is colonized by cold-water coral reefs. Large deep-water carbonate mounds formed by the interaction between reef-building cold-water corals and sedimentary processes are only found on the margins from Ireland to the Gulf of Cadiz (Roberts et

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al., 2006; Wheeler et al., 2007). Azooxanthellate reef-building corals are mainly *Lophelia pertusa* and to a minor degree *Madrepora oculata* (Freiwald, 2002; Wienberg et al., 2008). Cold-water coral reefs are carbonate factories (Dorschel et al., 2007a; Titschack et al., 2009) occurring in water depths between 500 and 1000 m (Foubert et al., 2005; Wheeler et al., 2007) and reaching heights of up to 350 m and can be several kilometres in diameter.

In general, cold-water corals tolerate a wide range of environmental factors such as temperatures of 4–12 °C (Freiwald, 2002) and salinity values of 32–36 psu in the North Atlantic. In the Mediterranean Sea they thrive in waters with temperatures up to 14 °C and salinities up to 38.8 psu (Freiwald et al., 2009). Furthermore, cold-water corals tolerate values of dissolved oxygen ranging from 3.75 to 6.65 ml/l and grow in a large range of water depths (Dullo et al., 2008). The shallowest living colonies are found in 40 m water depth in the Trondheimsfjord, Norway (Fosså et al., 2005), while the deepest colonies are reported from 3273 m on the New England Seamount Chain in the northwest Atlantic (Freiwald et al., 2004). A controlling parameter for living cold-water coral reef distribution is the density of seawater. Dullo et al. (2008) demonstrated a relationship between the distribution of cold-water coral reefs and the hydrography on the Celtic and Norwegian Margin. Living reef ecosystems of cold-water corals in the Northeast Atlantic seem to occur within a density range of sigma-theta $(\sigma_\Theta) = 27.35$ to 27.65 kg m^{-3}. However, cold-water corals in the Mediterranean Sea seem to tolerate a different seawater density value of 29.1 kg m^{-3} (Freiwald et al., 2009).

Many studies have investigated the carbonate mounds in the North Atlantic during the last decade (De Mol et al., 2002, 2007; Dorschel et al., 2005, 2007a; Eisele et al., 2008; Hovland et al., 1994; Huvenne et al., 2005, 2007; Mienis et al., 2006, 2007, 2009; Rüggeberg et al., 2005, 2007; van Weering et al., 2003; Wheeler et al., 2005a,b, 2007; Wienberg et al., 2008), but the initiation and start-up phase of these structures have been only recently studied (Foubert and Henriet, 2009; Huvenne et al., 2009; Kano et al., 2007; Louwye et al., 2007; Sakai et al., 2009; Titschack et al., 2009).

It is presently known that cold-water coral mound growth is initiated by the correct interplay of all necessary environmental condition and that mound growth occurred in cycles along the European continental margin. Previous works (e.g., Roberts et al, 2006; Rüggeberg et al., 2007) showed that mound development generally occurred during interglacials, whereas mounds are inactive during glacial times. During warmer periods (interglacial) stronger currents supply more nutrients making cold-water coral growth favourable. During glacial times weak currents, decreased nutrient supply and enhanced sedimentation rates do not support coral growth (Dorschel et al., 2005, 2007a; Roberts et al., 2006; Rüggeberg et al., 2005, 2007). However, the most-recent models are based on short gravity cores from the Porcupine Seabight (PSB), and hence span only the last interglacial/glacial cycles. The studies of Foubert and Henriet (2009), Kano et al. (2007) and Titschack et al. (2009) unveiled for the first time the full duration of a cyclic mound build-up at the Challenger Mound.

This study focuses on the stable isotopic signature (δ^{18}O and δ^{13}C) of planktonic (*Globigerina bulloides*) and benthic foraminifera (*Fontobtia wuellerstorfi, Discanomalina coronata, Lobatula lobatula, Lobatula antarctica* and *Planulina ariminensis*), and sediment grain size data (mean sortable silt) of sediments from IODP Exp. 307 Site U1317C. Our aim is to quantitatively reconstruct at a high-resolution the paleoenvironmental and paleoecological setting favourable for initial coral settling and development on the Challenger Mound.

1.1. Regional setting and hydrography

The Porcupine Seabight is an amphitheatre-shaped embayment. It is about 150 km long, 65 km wide and located southwest of Ireland in the North Atlantic (Fig. 1). To the southwest the Porcupine Seabight passes into the Porcupine Abyssal Plain at a depth of 3000 m, while it is limited by the Slyne Ridge in the north at a depth of 250 m. More than 1600 carbonate mounds possibly occur in the region (Foubert et al., 2005). Five main mound provinces are present in the Porcupine Seabight (Fig. 1): (1) the Magellan Mound Province in the north, (2) the Hovland Mound Province further south, (3) the recently discovered Viking Mound Province southeast of the Hovland Mound Province, (4) the Belgica Mound Province including the Challenger Mound at its eastern margin, and (5) the Enya Mound Province south to southeast of the Belgica Mound Province (De Cock, 2005; De Mol et al., 2002; Hovland et al., 1994; Van Rooij et al., 2007). The Belgica Mound Province is about 45 km long. Here De Mol et al. (2002) described 66 conical mounds, occurring in water depths of 550 to 1025 m.

A detail review of the present-day oceanographic setting of the Porcupine Seabight in relation to the carbonate mounds is given by White et al. (2005). A general northward along-slope current system originates at the Iberian Margin and flows along the East-Atlantic margin into the Norwegian Sea. The most important water masses at intermediate depths are the Eastern North Atlantic Water (ENAW) and the Mediterranean Outflow Water (MOW). The Eastern North Atlantic Water reaches a water depth down to 800 m and is underlain by the MOW. At 950 m water depth the MOW is characterized by an oxygen minimum and a salinity maximum (Pollard et al., 1996). The Labrador Sea Water and the Norwegian Sea Deep Water are reported to occur at greater depths below the MOW (Rice et al., 1991). In the Belgica Mound Province the strong hydrodynamic regime is combined with the presence of internal waves and tides at the boundary between ENAW and MOW, hence underlining the unique setting of the Belgica Mound Province (De Mol et al., 2002; Rice et al., 1991).

The Challenger Mound, located on the eastern margin of the Porcupine Seabight, is a 155 m high carbonate mound covered with dead cold-water coral fragments (Foubert et al., 2007). It was drilled during IODP Expedition 307 at Site U1317 (Fig. 1, 51 °22.8′ N, 11°43.1′ W; 781–815 m water depth) (Williams et al., 2006). The up to 155 m long coral-bearing sedimentary successions have great potential to shed light on the processes triggering the colonization of cold-water corals on the mounds and the driving mechanisms of mound growth itself. A thick sedimentary cover is documented on the upslope flank of the mound, while a thinner one occurs on the downslope flank. The Challenger Mound is characterized by different growth phases and an erosional boundary overlying glauconitic and silty sandstone drift deposits at the mound base (Expedition Scientists, 2005; Foubert and Henriet, 2009; Kano et al., 2007). Sediments below the base of the Challenger Mound are of middle Miocene age (14.78–15.16 Ma; Louwye et al., 2007), whereas the mound base is about 2.6 Ma old (Foubert and Henriet, 2009; Kano et al., 2007). A second major hiatus identified at 1.7 Ma suggests that during this time the mound was not active. The overlaying sediments are dated at 1 Ma and indicate a possible re-activation of the mound (Foubert and Henriet, 2009; Kano et al., 2007).

2. Material and methods

This study is based on sediments from Hole U1317C recovered during IODP Expedition 307 with R/V *Joides Resolution* in 2005 (Expedition 307 Scientists, 2006). In order to obtain undamaged half cores, they were frozen before splitting (Dorschel et al., 2005; Foubert et al., 2007).

At 147.95 msbf (metres below seafloor) sediments from core U1317C are characterized by an unconformity, marked by a sharp colour change from the grey, coral bearing sediments to the greenish-grey underlying unit. Samples were taken every 10 cm in the interval between 141 and 151 mbsf just above and below the mound base using 10 cm^3 syringes. All samples were dried at 50 °C, weighed and then wet sieved through a 63-µm sieve. The suspended fine fraction (<63 µm) was collected in 3-litre jars for fine fraction analysis. The coarse fraction was again oven dried, weighed and dry sieved at 125 µm and 250 µm to obtain the correspondent size fraction.
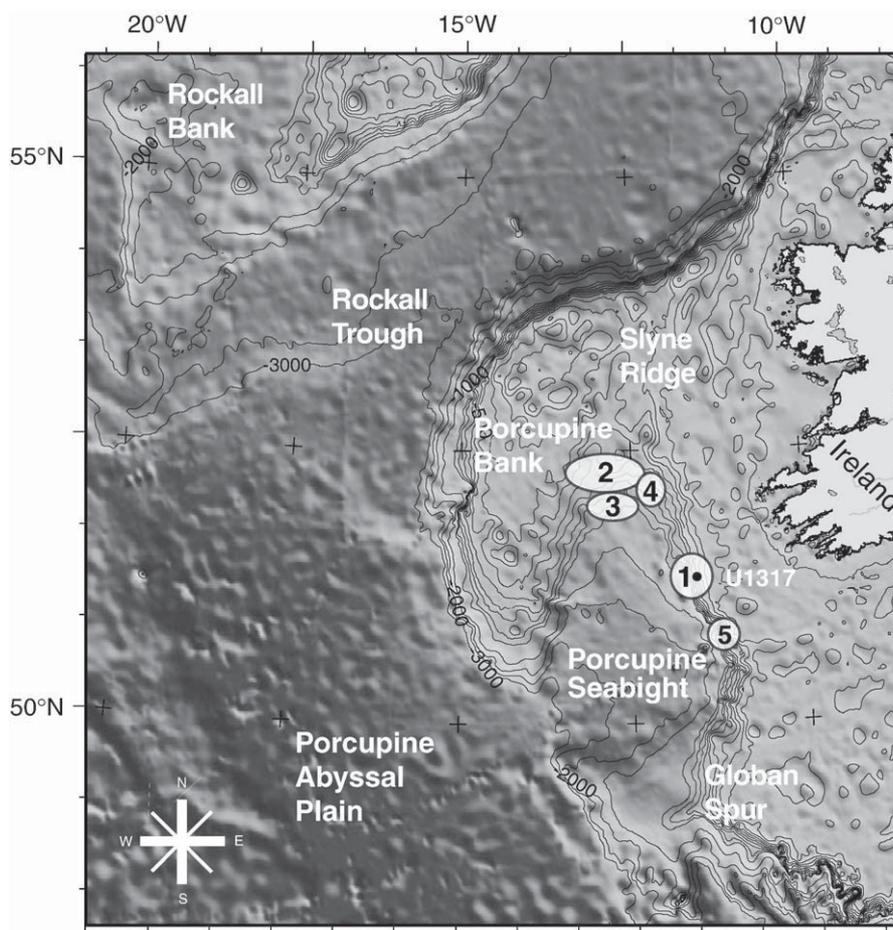
Fig. 1. Map showing the bathymetry of the North-East Atlantic with the Irish and Celtic Margin. The studied sediment drill core (Site U1317C) is indicated within the Belgica Mound Province (1). Other mound provinces are the Magellan (2), the Hovland (3), the Viking (4), and the Enya Mound Province (5). Datasets for the map are based on the ETOPO5 digital elevation file (http://www.ngdc.noaa.gov/).

2.1. Grain size analyses

Grain-size distribution of the fine fraction (<63 μm) was determined from the collected fine fraction with a Micromeritic Sedigraph 5100. This device measures the concentration of sediment in suspension by the attenuation of an X-ray beam. Water in the samples was replaced by a sodium polyphosphate solution (0.05%) to avoid flocculation of particles. After homogenisation on a rotating carousel (at least 12 h), the samples were placed in the ultra-sounds for 10 s. Afterwards they were analysed with a density setting of calcite ($2.71 \, g \, cm^{-3}$) at a constant water temperature of 35 °C and with an analysis range from 1 to 63 μm. Cumulative and mass frequency data output was used to calculate mean silt (0–63 μm) and mean sortable silt (10–63 μm) distributions, along with size frequency distributions. Finally raw data were converted into weight percentages (wt.%).

2.2. Stable isotope analyses

Stable oxygen and carbon isotope analyses ($\delta^{18}O$ and $\delta^{13}C$) were carried out on well-preserved and clean foraminifera specimens. The benthic species *Fontbotia wuellerstorfi* (3 specimens), *Planulina ariminensis* (5), *Lobatula antarctica* (3), *Lobatula lobatula* (5), and *Discanomalina coronata* (3), were picked from the size fraction larger than 250 μm and the planktonic species *Globigerina bulloides* (15) was picked from the >125 μm size fraction (Plate 1). Oxygen and carbon isotopes were measured with a Finnigan 252 mass spectrometer with a Kiel CARBO device at IFM-GEOMAR in Kiel for the species *L. lobatula* and *G. bulloides* and due to technical problems also at the Isotope Laboratory of the Institute of Geology and Mineralogy at the University of Erlangen for the species *F. wuellerstorfi*, *D. coronata*, *L. antartica* and *P. ariminensis*. Approximately 0.1 mg of $CaCO_3$ was measured for each sample. Reproducibility was ± 0.045‰ for $\delta^{18}O$ and ± 0.012‰ for $\delta^{13}C$. Isotope ratios are presented relative to the PeeDee Belemnite (PDB) standard based on calibration with National Bureau of Standards (NBS).

Bottom-water temperatures were calculated using the equation of Shackleton (1974) for benthic foraminifera: $T (°C) = 16.9 - 4.38 (\delta^{18}O_c - \delta^{18}O_w) + 0.10^* (\delta^{18}O_c - \delta^{18}O_w)^2$ using a $\delta^{18}O_w$ of 0‰ for the Early Pleistocene/Late Pliocene and -0.25‰ for the Miocene (Zachos et al., 2001).

Sea Surface Temperature (SST) was calculated using the $\delta^{18}O_c$ values of the planktonic species *Globigerina bulloides*. The equation of Erez and Luz (1983) was applied using a $\delta^{18}O_w$ of 0‰ for the Early Pleistocene/Late Pliocene and -0.25‰ for the Miocene (Zachos et al. 2001) as follows: $T (°C) = 16.998 - 4.52 (\delta^{18}O_c - \delta^{18}O_w) + 0.028 (\delta^{18}O_c - \delta^{18}O_w)^2$.

3. Results

3.1. Stable oxygen and carbon isotopes in foraminifera

Variations in the oxygen and carbon isotope records of benthic foraminifera are shown in Figs. 2 and 3. Oxygen isotope values of benthic foraminifera have a similar mean value of ~2‰ for all species. High frequency variations characterize the patterns at small scale. *Fontbotia wuellerstorfi* displays a variability of ~0.8‰ in the investigated interval above the mound base <149 mbsf. Oxygen isotope values of the other species vary by ~0.8‰ (*Lobatula antartica*), ~1‰

(*Lobatula lobatula*), ~1.3‰ (*Discanomalina coronata*), and ~1.1‰ (*Planulina ariminensis*). Downcore the $\delta^{13}C$ values have a different pattern with respect to $\delta^{18}O$.

The mean $\delta^{13}C$ values are ~0.25‰ for *Fontbotia wuellerstorfi*, ~0.6‰ for *Discanomalina coronata* and *Planulina ariminensis*, ~0.1‰ for *Lobatula antartica*, and ~0.15‰ for *Lobatula lobatula*. The variability of $\delta^{13}C$ values is much larger than for the $\delta^{18}O$. Carbon isotope values of *F. wuellerstorfi* vary by ~1.8‰ and are comparable to *P. ariminensis*. *Lobatula antartica* and *L. lobatula* show slightly higher variability of ~2‰ and ~2.2‰, respectively. Overall, $\delta^{13}C$ variations of *D. coronata* of ~2.5‰ are the highest of all the species investigated.

Oxygen and carbon isotope ratios of *Globigerina bulloides* change rapidly at ~148 mbsf (Figs. 2–4) with an $\delta^{18}O$ increase by 2.5‰ and a $\delta^{13}C$ decrease by about 2‰. Similar $\delta^{18}O$ values of *Lobatula lobatula* increase by about 2‰, and the $\delta^{13}C$ values decrease by 1‰ (Figs. 2–4).

Temperature reconstructions were performed following Shackleton (1974) for the benthic and Erez and Luz (1983) for the planktonic species. Both records display a distinct shift at the mound base at ~148 mbsf. Temperature estimates decrease from 21° to 10° for the sea surface and from 17° to 9 °C for the deep waters between the Mid Miocene and the mound initiation in the Pliocene (Fig. 4). Over the whole record Sea Surface Temperature (SST) values estimated from *Globigerina bulloides* display a higher variability than Bottom Water Temperature (BWT) estimates (3–4 °C). The same pattern is visible in the $\delta^{13}C$ records, where the variability of planktonic $\delta^{13}C$ values is about 2‰ larger than the benthic.

3.2. Grain size analyses

The sortable silt fraction of carbonate free sediment is sensitive to hydrodynamic processes and can be used to estimate variations in paleocurrent intensities (McCave et al., 1995). In core U1317C, mean sortable silt is characterized by two different patterns: (1) values in the upper interval (~141 to ~148 mbsf depth) vary from about 21 to 27 μm, whereas (2) values in the lower interval (~148 to ~151 mbsf) only vary from 18 to 22 μm (Fig. 4). The calculated means for each interval are 23.66 (±2.76) μm for the upper part and 19.74 (±1.90) μm for the lower part.

4. Discussion

4.1. Species reliability

Stable isotope measurements on foraminifera are an important tool in paleoceanographic studies and are routinely used for paleoenvironmental reconstructions. In general oxygen isotope ratio of foraminiferal calcite reflects the $\delta^{18}O$ value of seawater ($\delta^{18}O_w$), which varies with the global ice volume as a function of seawater salinity (Lynch-Stieglitz et al., 1999). Additionally, the foraminiferal calcite exhibits $\delta^{18}O$ fractionation dependent on temperature and therefore $\delta^{18}O$ is used as a proxy for temperature.

The $\delta^{13}C$ value of foraminiferal calcite tests is primarily a function of the $\delta^{13}C$ value of dissolved inorganic carbon (DIC) in the seawater (Emiliani, 1955; McCorkle et al., 1990; O'Neil et al., 1969) and has been used in studies as a proxy for productivity and to trace water masses. Further factors influencing the $\delta^{13}C$ are interspecies offsets (Duplessy et al., 1970; Shackleton, 1974) attributed to the microhabitat preferences of benthic foraminifera (deep infaunal, epifaunal, elevated substrates or even fauna-attached; Mackensen et al., 2000; McCorkle et al., 1990), vital effects (species-specific metabolic

variation), and any potential diagenetic overprint. Hence, prior to any interpretation of paleoceanographic relevance we compare isotopic data of multiple species for their reliability of isotopic information.

Not all of the benthic foraminiferal species studied here occur continuously in middle Miocene sediments (below the mound base). Only *Lobatula lobatula* is abundant throughout the entire analysed section, while *Discanomalina coronata*, *Lobatula antartica*, *Fontbotia wuellerstorfi*, *Planulina ariminensis* occur from 148 mbsf upward. Lutze and Thiel (1987) have shown that *F. wuellerstorfi* only colonize elevated epibenthic microhabitats up to 14 cm above the seafloor. Hence the $\delta^{13}C$ signal of this species can be used as a recorder of the ambient bottom water properties (Graham et al., 1981; McCorkle et al., 1990; Zahn et al., 1997). Although measured at different laboratories, the oxygen and carbon isotope values of the benthic foraminifera *F. wuellerstorfi*, *L. antarctica*, *P. ariminensis* and *L. lobatula* are similar and therefore useful for deep-water paleoceanographic reconstructions at this location. Where present, all the investigated species display stable oxygen and carbon isotope values that are consistent with those of *F. wuellerstorfi*, indicating that they respond to similar environmental conditions and are influenced by similar bottom waters. Only $\delta^{13}C$ values of *D. coronata* are significantly offset from other species by around −2‰. This species is generally associated with cold-water coral ecosystems (Margreth et al., 2009) and lives attached on elevated substrate like a dead coral framework. The $\delta^{13}C$ offset of *D. coronata* with respect to the other species most likely results from a strong vital effect, also known for some other benthic species (Wilson-Finelli et al., 1998). These species have a strong metabolic effect causing low $\delta^{13}C$ values, whereas kinetic effects result in even more depleted $\delta^{13}C$ values (Mackensen and Bickert 1999). As *G. bulloides* and *L. lobatula* provide a continuous downcore record across the base of Challenger Mound, we only use their $\delta^{18}O$ and $\delta^{13}C$ records for downcore paleoenvironmental interpretation.

4.2. Paleoenvironmental conditions during initiation and early development of Challenger Mound

Carbonate mound growth on the European margin is generally controlled by environmental conditions. In particular the first settlement of cold-water coral larvae is thought to result from the introduction of the MOW into the PSB (De Mol et al. 2002; Freiwald 2002). Development of a single mound occurs from localised coral larvae settlements, supporting the build-up of a giant carbonate mound (Huvenne et al., 2009). Further development is driven by environmental changes following glacial−interglacial cycles (Rüggeberg et al., 2007). Initial growth of the Challenger Mound occurred at the beginning of the Northern Hemisphere Glaciation at ~2.6 Ma (Kano et al., 2007) and mound growth within the entire PSB probably occurred at the same time (De Mol et al., 2002).

Our study provides insight into the paleoenvironmental setting during the start-up phase and initiation of Challenger Mound. In the following discussion we concentrate on (Section 4.2.1) the setting of the middle Miocene recorded in the sediments below the mound base around 15 Ma, (Section 4.2.2) the mound initiation at ~2.6 Ma and first episode of mound growth at the mound base after a hiatus of ~12 Myr, and (Section 4.2.3) on the climatic and oceanographic variabilities indicating variations in early mound development in comparison to present-day settings.

Plate 1. List of genera and species used in this study. The families are listed in taxonomic order following Loeblich and Tappan (1988), genera and species are listed in alphabetical order. 1 a–c cf. *Lobatula antarctica* (Saidova, 1975), sample 1317 C, 17-1, 54–55 cm. 2 a–c *Fontbotia wuellerstorfi* (Schwager, 1866), sample 1317c C, 17-1, 44–45 cm, 3 a–c *Lobatula lobatula* (Walker and Jacob, 1798), sample 1317 C, 17-4, 90–91 cm, 4 a–c. *Planulina ariminensis* (d'Orbigny, 1826), sample 1317 C, 17-3, 27–28 cm. 5 a–c. *Discanomalina coronata* (Parker and Jones, 1857), sample 1317 C, 17-4, 90–91 cm. 6 a–c. *Globigerina bulloides* (d'Orbigny, 1826), sample 1317 C, 17-1, 24–25 cm. a = umbilical views, b = side views, c = spiral views.
Data from Saidova (1975), Schwager, (1866), Walker and Jacob (1798), d'Orbigny (1826), Parker and Jones (1857), and d'Orbigny (1826).
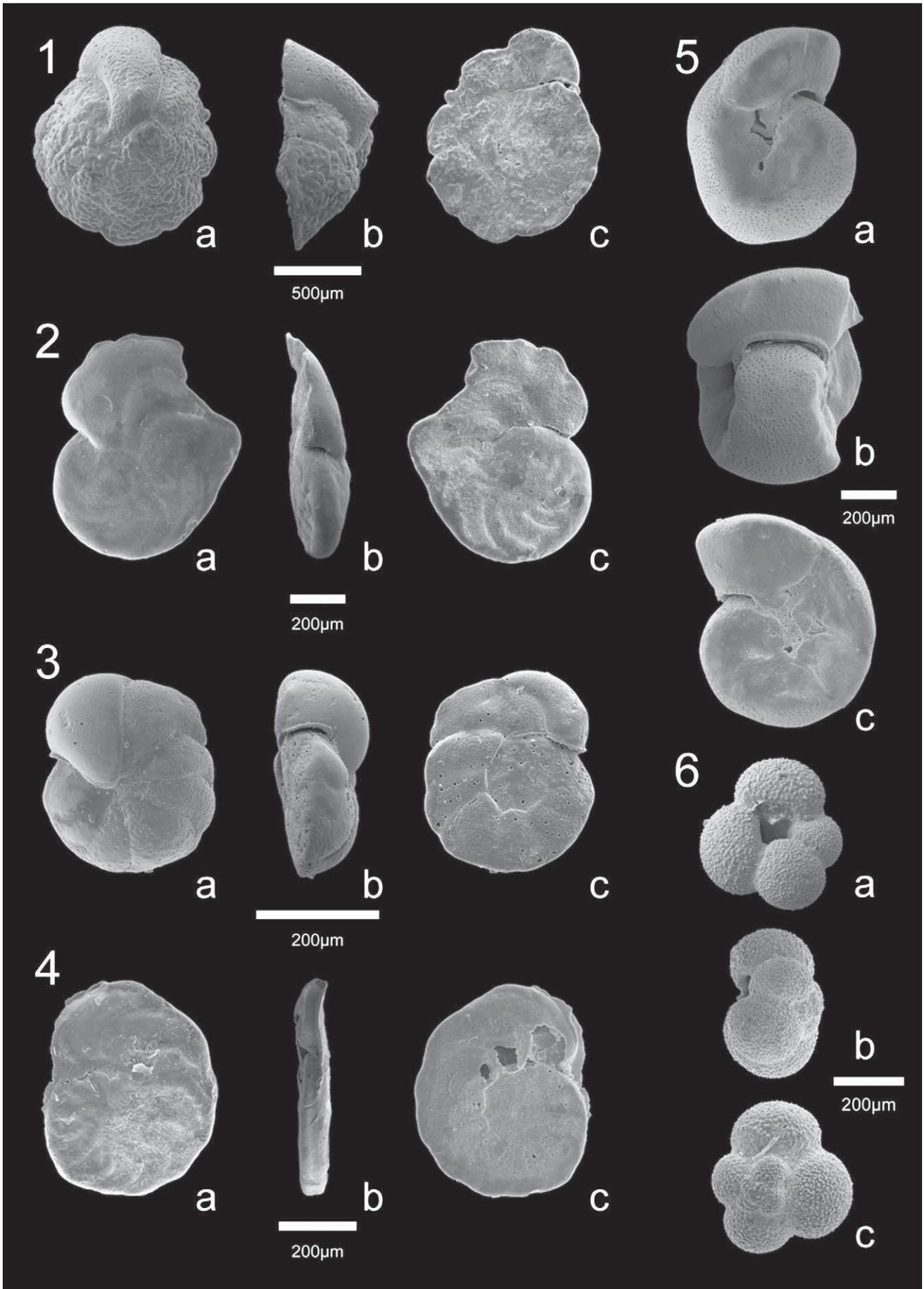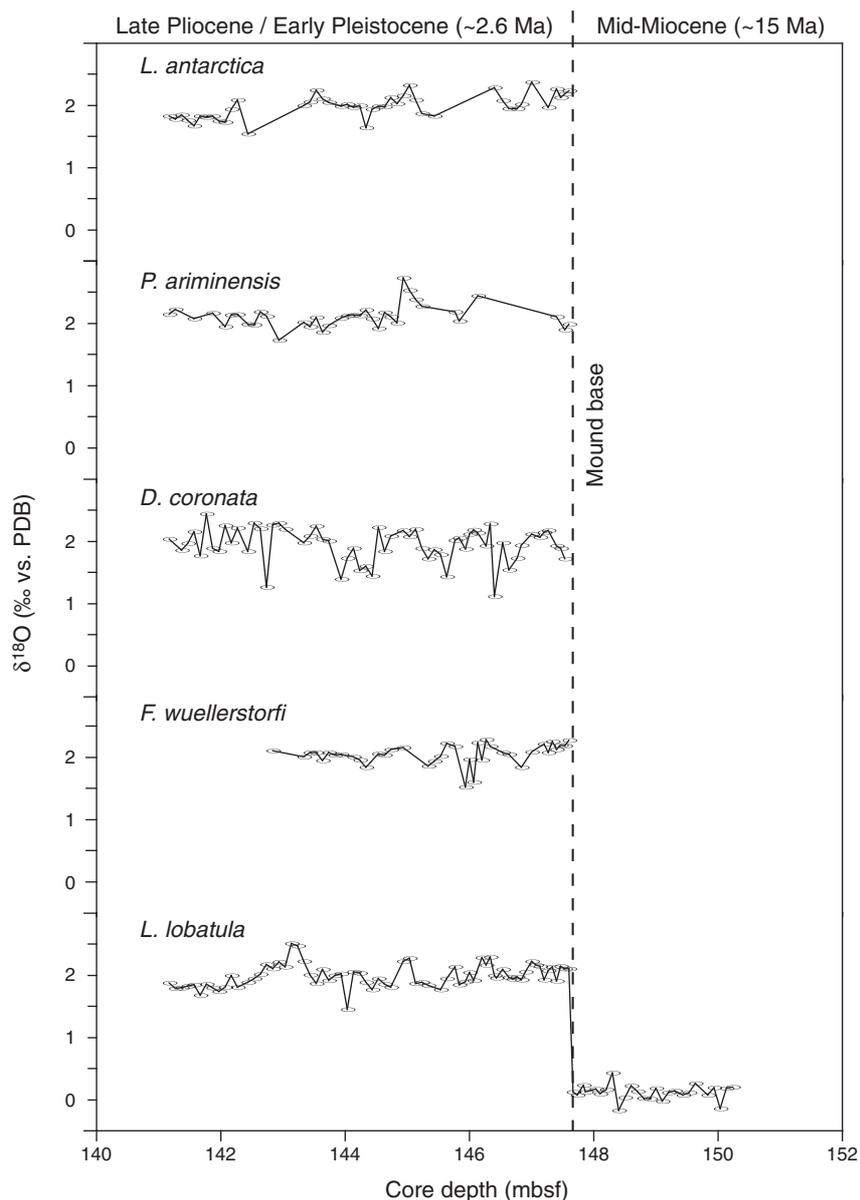
Plate I.

Fig. 2. δ^{18}O values (in ‰ vs. PDB) of benthic foraminifera *Lobatula lobatula*, *Fontbotia wuellerstorfi*, *Discanomalina coronata*, *Planulina ariminensis*, and *Lobatula antartica*.

4.2.1. Below the mound base

The period of the middle Miocene between 14 and 17 Ma was the warmest of the Neogene and is known as the Miocene Climatic Optimum (MCO) (Zachos et al., 2001). Several studies of the middle Miocene indicate warm SST ranging between 15° and 21 °C at a latitude of around 50°N (e.g., Nikolaev et al., 1998; Pagani et al., 1999; Savin, 1977; Shevenell et al., 2004). Our middle Miocene SST data from IODP Site U1317C based on *Globigerina bulloides* are consistent with the reported SSTs varying between 16 and 22 °C (Fig. 4). The reconstructed BWT of ~15 °C also indicates a very warm intermediate water mass for the MCO at ca. 800 m water depth. Such warm BWT or light benthic stable oxygen isotope values are also reported from the low latitude Atlantic during the latest early Miocene (15–17 Ma; Nikolaev et al., 1998) and from deeper sites in the central North Atlantic (Pagani et al., 1999; Wright et al., 1992). During this time a weak circulation pattern occurred in the North Atlantic caused by a reduction of the Northern Component Water formation (Miller and Fairbanks, 1983; Zachos et al., 2001), while warm and saline

intermediate waters of southern origin prevailed (Flower and Kennett, 1994; Wright et al., 1992). Relatively high planktonic and benthic foraminiferal δ^{13}C values between 1 and 2‰ support the origin of this warm and saline, possibly Tethyan intermediate water mass (Fig. 4).

High δ^{13}C values during the early to middle Miocene are also considered to be the result of large-scale changes in organic carbon deposition relative to carbonate sedimentation, the so-called "Monterrey Carbon Excursion" (Vincent and Killingley, 1985). Intervals of high organic carbon accumulation in marginal basin sediments are marked by distinct δ^{13}C maxima in the benthic record between 17 and 15 Ma with values of 1.5–2‰ (Flower and Kennett, 1994; Wright et al., 1992). Remarkable high planktonic and benthic δ^{13}C signals may reflect this pivotal point in Cenozoic climatic evolution at Hole U1317C (Fig. 4).

The MCO-following cooling phase is not documented in the cores of IODP Site 1317. Long hiatuses of more than 3 Myr were common in the late Miocene to Pliocene NE Atlantic and represent major erosional (and/or nondepositional) events in low-productivity
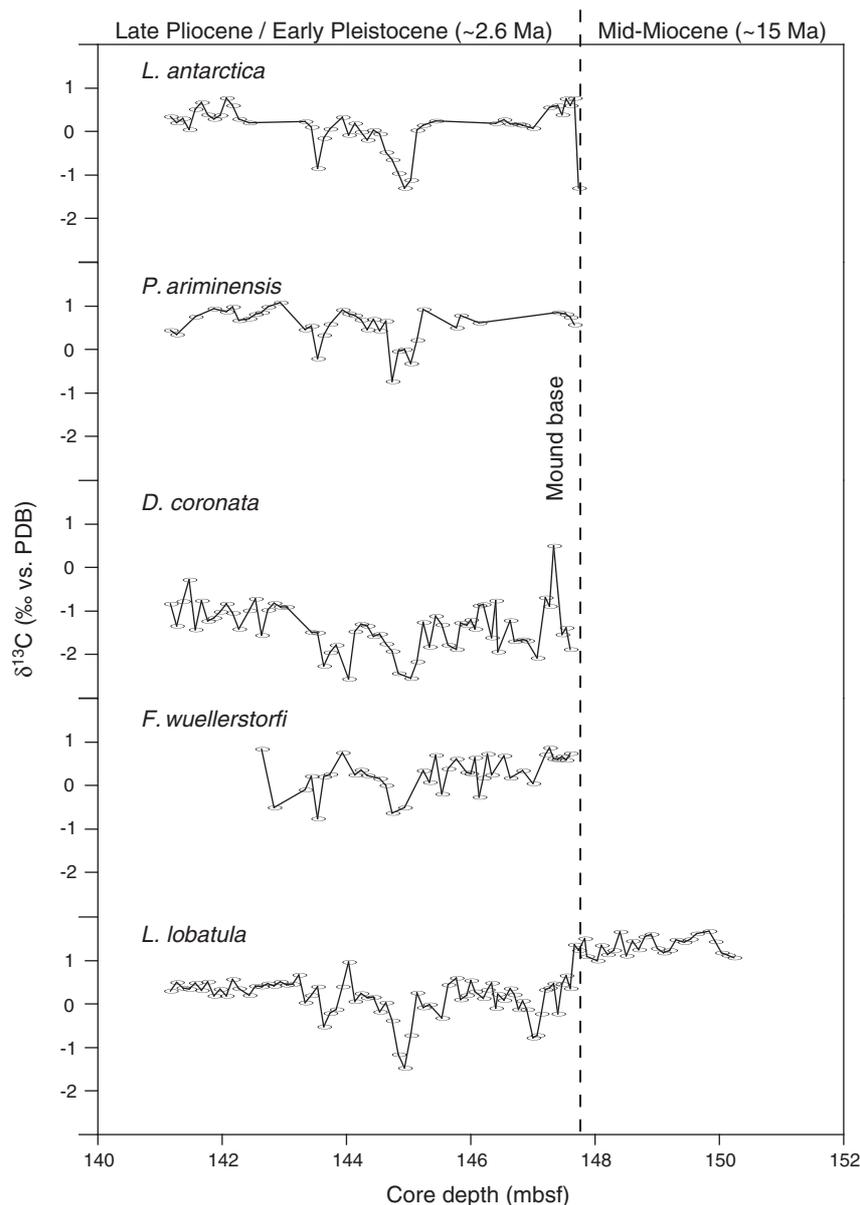
Fig. 3. δ^{13}C values (in ‰ vs. PDB) of benthic foraminifera *Lobatula lobatula*, *Fontbotia wuellerstorfi*, *Discanomalina coronata*, *Planulina ariminensis*, and *Lobatula antartica*.

regions or paths of strong bottom currents (Keller and Barron, 1983; Pearson and Jenkins, 1986). According to these authors Neogene hiatuses (NH) occurred at 13.5 to 12.5 Ma (NH3), 12–11 Ma (NH4), 10–9 Ma (NH5), 7.5–6.2 Ma (NH6), and 5.2–4.7 Ma (NH7) documented in different DSDP cores along the European continental margin at water depths between 1600 and 4500 m. All these hiatuses at intermediate depth developed during cold periods and may therefore rather be attributed to changes in the hydrodynamic regime at intermediate waters as suggested by Louwye et al. (2007) than to low productivity. This is comparable to dynamics described for the carbonate mound settings during glacial–interglacial cycles of the late Pleistocene (Dorschel et al., 2005; Rüggeberg et al., 2007; Van Rooij et al., 2007). The sum of hiatuses NH3–NH7 (13.5–4.7 Ma) give a reasonable explanation for the lack of sediments spanning ~12 Myr in core 1317C and the erosional unconformity, on which Challenger Mound started to develop at ~2.6 Ma (Kano et al., 2007).

4.2.2. Mound initiation and first episode of mound growth

The initiation of coral mound growth is indicated by the sharp increase in planktonic (benthic) δ^{18}O values of ~3 (2)‰, 4 µm increase

of mean sortable silt and δ^{13}C decrease of ~1.5–2‰, respectively (Fig. 4). Compared to δ^{18}O values of the same species from core 1317E (Sakai et al., 2009), our δ^{18}O values of *Globigerina bulloides* are ~0.5‰ heavier. This offset cannot be simply explained by differences in the resolution of data. Sakai et al. (2009) defined the beginning of mound growth to Marine Isotope Stage (MIS) 92 at ~2.24 Ma, which is a cold or early glacial period. However, Foubert and Henriet (2009), Huvenne et al. (2009) and Kano et al. (2007) argue for an earlier onset of mound initiation in core 1317E. A diachronous growth of Challenger Mound is the best explanation for different initiation dates between the cores of Site U1317. Foubert and Henriet (2009) were able to show this diachronous mound growth using spectral analyses on geophysical and geochemical parameters. They demonstrated that the nucleation of Challenger Mound started before 2.58 Ma close to Hole U1317E and later at Hole U1317C (<2.58 Ma), reflecting therefore not the entire mound history in the latter hole.

Despite the different mound start-up phases in Holes U1317E and U1317C, the heavier planktonic oxygen isotope values of ~2‰ at site 1317C (Fig. 4) point to mound development during an early Pleistocene glacial period. The Δ^{18}O$_{plankt}$ between glacial and interglacial phases
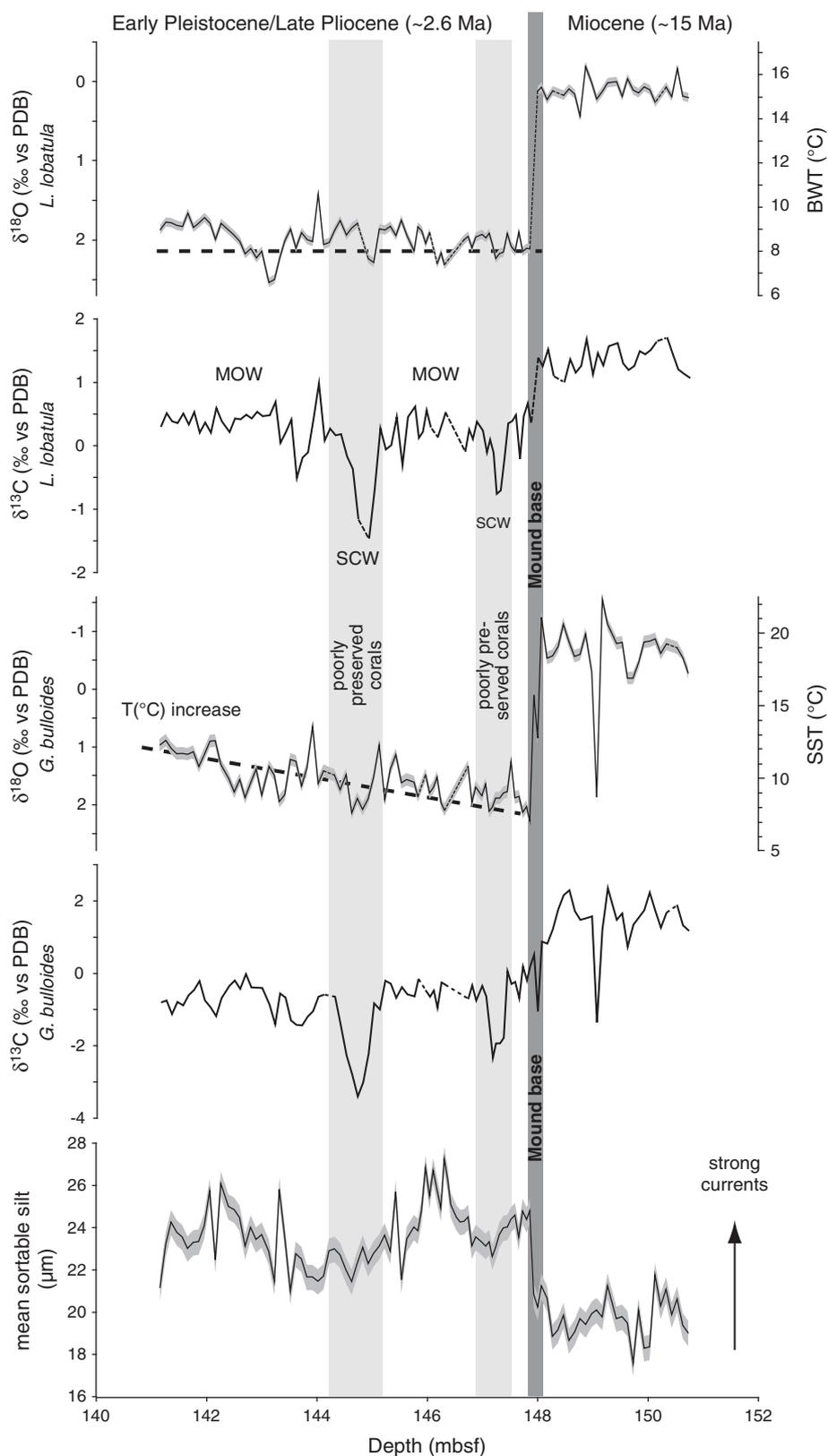
Fig. 4. High-resolution records of IODP Site U1317C, 141–151 mbsf. Mean sortable silt (in μm), stable oxygen and carbon isotopes (in ‰ vs. PDB) from *Lobatula lobatula* and *Globigerina bulloides*. Stable oxygen isotope values were converted into Sea Surface Temperatures (SST in °C) and Bottom Water Temperatures (BWT in °C). Grey envelope indicates error of reconstructed temperatures. Dotted line between 141 and 148 mbsf at benthic δ^{18}O indicates mean Mediterranean Outflow Water (MOW) temperatures of ~8 °C (from Khélifi et al., 2009). Light grey bars indicate sections of poorly preserved corals coherent with possible occurrence of Southern Component Water (SCW).

~2.5 Ma ago is ca. 0.7–1.3‰ (Lisiecki and Raymo, 2005). The small-scale variability in planktonic oxygen isotope data of core 1317C is in the same order of ~1‰ and the general trend form the mound initiation to the end

of the record at 141 mbsf as well (Fig. 4, see Section 4.2.3). Taking a constant sedimentation rate of 15 cm kyr^{-1} (max. 24 cm kyr^{-1}) without any hiatus into account (Kano et al, 2007), the investigated

section (141–148 mbsf) comprises only a part of a glacial–interglacial cycle with ~47 kyr (~30 kyr), respectively. Therefore, we assume that Challenger Mound initiation may have occurred in an early Pleistocene glacial phase, amplified by the fact that glacial periods were less extreme at that time than the more recent ones (Lisiecki and Raymo, 2005). This is in contrast with Huvenne et al. (2009), which relate the mound initiation to warmer conditions reconstructed from the characterization of the sedimentation mode and bottom current intensities. However, their planktonic foraminiferal assemblages display no clear difference within the sedimentary facies between 141 and 148 mbsf. This is also expressed in the relatively small variability of BWT of mean 8.56 ± 0.73 °C supporting small glacial–interglacial changes at that time and being still in the range of tolerated temperatures for the reef-building coral *Lophelia pertusa* (Freiwald, 2002).

Small variations in BWT are generally related to the interplay of different water masses or vertical movements of an intermediate water mass (see Section 4.2.3). During the initial glaciation of the northern hemisphere Mediterranean Outflow Water (MOW) dominated the oceanographic setting at intermediate water depths in the Porcupine Seabight (Khélifi et al., 2009). Accordingly, BWT from Site U1317C shows a striking similarity to Mg/Ca-based BWT of *Cibicides mundulus* from DSDP Site 548 (South of the Porcupine Seabight, 1250 m water depth) from 3.4 to 3.1 Ma (Khélifi et al., 2009). At this site BWT values around 8 °C indicate the influence of MOW in the North Atlantic.

It remains questionable why the build-up of Challenger Mound and possibly the other carbonate mounds in the Porcupine Seabight started several 100 ka after the intensification of MOW at 3.5–3.3 Ma (Khélifi et al., 2009), since MOW is assumed to be the main carrier of cold-water coral larvae from the Mediterranean Sea into the North Atlantic (De Mol et al., 2002). Probably vertical movements and a progressing shallowing of MOW after 3.3 Ma led to the initiation of coral development at the site of Challenger Mound. However, the inaccuracy of radiogenic Sr age determinations for that period should be kept in mind indicating a mound initiation between 2.329 and 3.614 Ma (Kano et al., 2007). Therefore the start-up of carbonate mound growth may be still linked to the introduction of MOW to the NE Atlantic.

4.2.3. Mound growth and short-term decline

The variability of stable isotope and grain size data indicates variations in the paleoenvironment during the early development of Challenger Mound. Bottom water temperatures remain relatively stable around 9 °C while SST shows an increase of ~3 °C indicating a shift from an early glacial phase (~144–148 mbsf) to an interglacial period (~141–144 mbsf) with comparable values as reported by Hooper and Funnell (1986) from Hole 552A between 2.4 and 2.6 Ma (Fig. 4).

Paleocurrent reconstructions from sortable silt analyses present a distinct increase of 4 μm, respectively, indicating a stronger current regime in the early Pleistocene compared to the Mid-Miocene (Fig. 4). Small-scaled variations in current velocities describe the early mound development with phases of reduced currents around 147 mbsf and between 144 and 145 mbsf, and intensified currents at the mound base, around 146 mbsf and between 142 and 143 mbsf. Schönfeld and Zahn (2000) described a similar cyclic pattern with ~5 μm amplitude at the Portuguese margin and relate it to changes of current intensities and short-term vertical movements of MOW within one glacial–interglacial cycle. After the Messinian salinity crisis (5.9–5.2 Ma; Soria et al., 2008) MOW underwent major changes in response to the successive glacial and interglacial stages, when sea level dropped by 50–120 m below to present level. Enhanced currents at greater depth and weaker current strength at shallower water depth characterized MOW during glacial periods. During terminations, when sea level rose and glaciers shrank, a MOW more similar to the recent one was established at shallower depth (Schönfeld and Zahn, 2000, Zahn et al.,

1997). Such a dynamic behaviour of MOW was probably not only restricted to recent glacial/interglacial cycles, but was also active, although less pronounced, in times of initial mound growth in the early Pleistocene.

Mean sortable silt values of 20 to 26 μm are comparable to modern and Holocene values of Propeller Mound (Hovland Mound Province; Rüggeberg et al., 2005), where residual currents of mean 2–5 cm s^{-1} occur (White, 2007). However, modern current intensities on top of Galway Mound, a mound 8 km north of Challenger Mound, reach 16 cm s^{-1}, whereas current speeds on its flanks appear to be slightly lower (Dorschel et al., 2007b). This suggests that current intensities in times of mound initiation were significantly weaker in the Belgica Mound Province than today. Huvenne et al. (2009) discussed a hydrodynamic regime characterized by an intermediate current strength optimal for cold-water corals to grow facilitating a fast, early mound development.

Trends in current intensity, productivity and sediment supply can be directly linked to episodes where non-deposition or erosion occurred. Titschack et al. (2009) showed that the early mound growth phase might also be disturbed by episodes of non-deposition and/or erosion. At Hole U1317C a gradual weakening of current strength and negative benthic δ^{13}C anomalies may be associated to changes in the bottom waters. Episodes of reduced currents at ~144.5 mbsf and at ~147.5 mbsf directly coincide with peak minima in the benthic and planktonic δ^{13}C values (Fig. 4). Extremely low planktonic δ^{13}C values of −2.5‰ and less can be related to upwelling of nutrient rich and δ^{13}C depleted waters during times of stronger advection (Naidu, 2004; Naidu and Niitsuma, 2004), whereas low benthic δ^{13}C values may reflect less ventilated bottom waters. The small temperature difference between bottom and surface waters at 144–148 mbsf may indicate a period of enhanced upwelling. This is supported by findings of Nikolaev et al. (1998) showing several vertical δ^{18}O profiles from the NE Atlantic (foraminiferal zone N21, 2–3 Ma) having ± constant oxygen isotope values of the upper 400 m of the water column. In the recent PSB, benthic δ^{13}C values of ~1‰ are recorded, while Holocene values are more depleted (0.3–0.5‰). During glacial periods with well-ventilated water masses, benthic δ^{13}C values were heavier with ~1.5‰ (Peck et al., 2007). Therefore, mean benthic δ^{13}C values of ~0.5‰ probably reflect relatively poorly ventilated bottom waters compared to the glacially enriched values resulting from the nutrient-depleted MOW mixing with ENAW (Frank et al., 2004).

The drop to extremely depleted δ^{13}C values at ~144.5 mbsf and at ~147.5 mbsf coincides with poorly preserved and partly dissolved coral skeletons (Fig. 4). What caused these extreme δ^{13}C decreases? Peck et al. (2007) also found brief episodes of depleted benthic δ^{13}C values during the last glacial/interglacial cycles in the Porcupine Seabight. They related these depleted δ^{13}C values to decreases in the ventilation of the Glacial North Atlantic Intermediate Water (GNAIW). These events are associated with melt water pulses from North Western European Ice Sheets (NWEIS). Becker et al. (2006) showed that the Late Pliocene/Early Pleistocene (~2.6 Ma) climatic system of the North Atlantic was influenced by similar short-term variations comparable to the recent oceanographic setting. Additionally, Thierens et al. (2010) showed a deposition of ice-rafted detritus already in the early stage of mound development. However, the maximum extent of NWEIS during the late Pliocene is not comparable to that of the last glaciations (Zachos et al., 2001). Therefore, other factors are required to explain the observed benthic δ^{13}C excursions.

Sarnthein et al. (1994) observed that the nutrient-rich δ^{13}C-depleted Southern Component Water (SCW) replaces MOW at intermediate depth during glacial periods. Decreased GNAIW formation and weak MOW flow during the early Pleistocene glacial periods may have lead to vertical redistribution of water masses along the European Margin (Peck et al., 2007; Zahn et al., 1987, 1997). This potentially led to the introduction of SCW into the Porcupine Seabight causing the observed depleted benthic δ^{13}C values. Therefore we
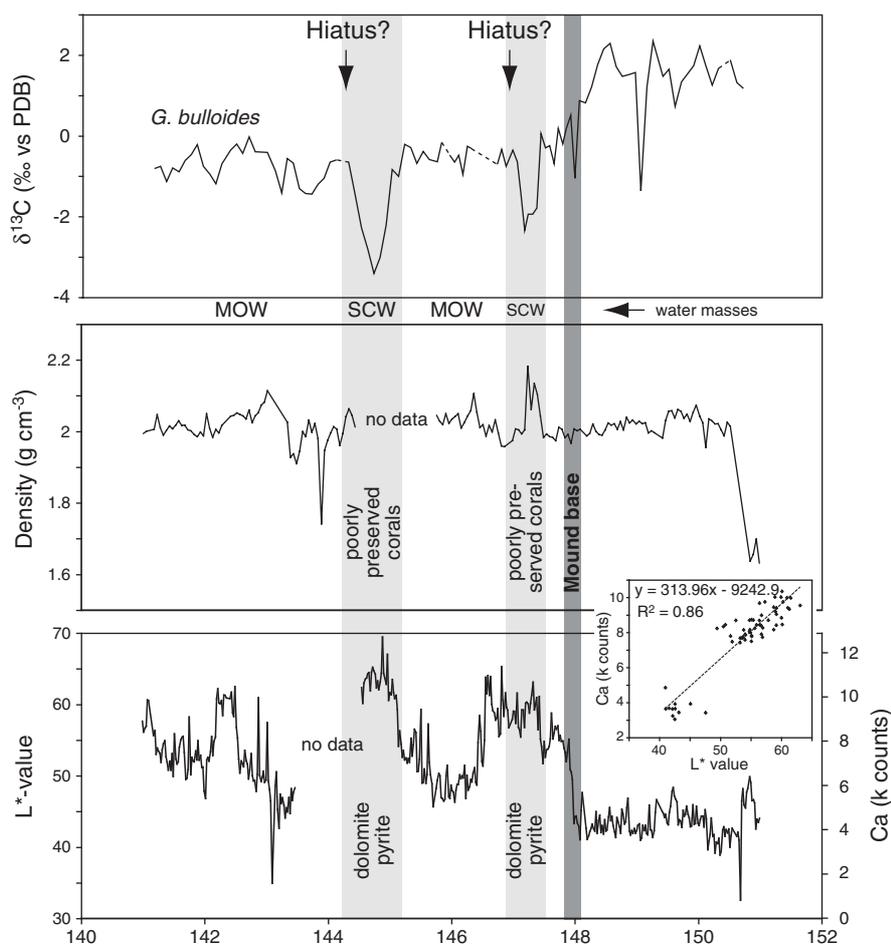
Fig. 5. Core logging data of sediment density (in g cm^{-3}), colour reflectance L*-value (in percent) and XRF Ca-counts (in k counts) of the investigated interval (Expedition 307 Scientists, 2006; Foubert and Henriet, 2009) are presented in relation to planktonic stable carbon isotopes (in ‰) and interpretations from Fig. 4. XY-plot indicates linear relation between L*-values and Ca-counts with correlation coefficient R^2 of 0.86. Sediment erosion after and early diagenetic (oxidation) processes within the grey intervals may have overprinted the sediments leading to dissolution of coral skeletons, the extreme δ^{13}C decreases, maxima in Ca-counts and sediment density, and the formation of pyrite and gypsum according to Pirlet et al. (2010).

conclude, that rather a glacial version of SCW influenced the oceanographic regime of the PSB during cold phases of the Early Pleistocene. In addition, occurrence of the cold-water benthic foraminiferal species *Lobatula antartica* underlines the appearance of a glacial SCW at these times (Figs. 2 and 3).

However, the introduction of SCW to intermediate depth in the PSB does not explain the even heavier drop of planktonic δ^{13}C values at the same time. Naidu and Niitsuma (2004) also reported δ^{13}C values of *Globigerina bulloides* about 1‰ lower than that of benthic foraminifera at a site in the Arabian Sea during the last glaciation. Pirlet et al. (2010) relate the occurrence of dissolved corals within a distinct sediment layer to the interplay of increased currents with possible sediment erosion and increased oxidation processes within the surface sediment layer. The result of this process is a distinct layer with a) dissolved coral skeletons, b) increased carbonate content and sediment density, c) the occurrence of gypsum crystals, d) first- and second-generation pyrite, and e) depleted δ^{13}C values of bulk sediment down to −5‰. Transferring these implications to the early development of Challenger Mound, both sections with low δ^{13}C values, and poorly preserved coral skeletons correlate well with higher sediment densities and Ca content (Fig. 5). A re-examination of the coarse sediment fraction from samples of dissolved coral layers showed that in these horizons pyrite and gypsum minerals are abundant compared to sediments with well-preserved coral skeletons. We therefore conclude that both, changes in the hydrography at surface and intermediate depth (SCW, upwelling, and MOW char-

acteristics) and early diagenetic (oxidation) processes induced by erosional events describe the sediments at the mound base of Site U1317C.

However, it is not possible to estimate the length of the hiatuses and how much material might have been eroded. A comparison of the δ^{18}O record of Challenger Mound with other records to determine Marine Isotope Stages (Sakai et al., 2009) must therefore be treated with caution due to the temporal gaps in the sedimentary record. A multi-proxy dataset in combination with early diagenetic structures as proposed by Pirlet et al. (2010) could resolve possible hiatuses through the whole Challenger Mound record.

5. Conclusion

This study provides detailed stable isotope records as well as grain size analyses from the base of IODP Site U1317C at Challenger Mound, Porcupine Seabight. These records encompass the sedimentary sequence from the Middle Miocene to the Early Pleistocene and reveal the presence of an extended hiatus between the middle Miocene and the Late Pliocene/Early Pleistocene. Mound growth coincided with the intensification of the Northern Hemisphere Glaciation, characterized by moderate glacial conditions that were still favourable for cold-water corals to grow. Temperatures calculated from benthic δ^{18}O are between 7 and 11 °C, consistent with the range of the known temperature tolerance (4–12 °C) for the reef forming cold-water coral *Lophelia pertusa*. Bottom current intensities are

characterized by a cyclic pattern that can be associated to vertical movements of Mediterranean Outflow Water (MOW) and its replacement by Southern Component Water (SCW). Peak events of these gradual movements are clearly observed in both planktonic and benthic $\delta^{13}C$ values indicating the influence of $\delta^{13}C$-depleted SCW in the Porcupine Seabight during that time. After these sluggish phases enhanced MOW flow replaced SCW and led to stronger current intensities and most probably sediment erosion on Challenger Mound. Erosion and early diagenetic (oxidation) processes overprinted the sediment layers as indicated by dissolved coral skeletons, the increase in Ca-content and sediment density, minimum $\delta^{13}C$ values, as well as the occurrence of gypsum and pyrite, implying a careful evaluation of original and overprinted geochemical signals.

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