

Evidence of early bottom water current flow after the Messinian Salinity Crisis in the Gulf of Cadiz



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ABSTRACT

Integrated Ocean Drilling Program (IODP) Expedition 339 cored multiple sites in the Gulf of Cadiz in order to study contourite deposition resulting from Mediterranean Outflow water (MOW). One Hole, U1387C, was cored to a depth of 865.85 m below seafloor (mbfs) with the goal of recovering the latest Miocene to Pliocene transition in order to evaluate the history of MOW immediately after the end of the Messinian Salinity Crisis. To understand this history, an accurate age model for the succession is needed, but is challenging to construct, because the Miocene–Pliocene boundary is not marked by a clear biostratigraphic event in the Atlantic and coring gaps occur within the recovered stratigraphic record. These limitations are overcome by combining a variety of chronostratigraphic datasets to construct an age-model that fits the currently available age indicators and demonstrates that coring in Hole U1387C did indeed recover the Miocene–Pliocene boundary at about 826 mbfs. This boundary is associated with a distinct and abrupt change in depositional environment. During the latest Messinian, hemipelagic sediments exhibiting precession-induced climate variability were deposited. These are overlain by Pliocene sediments deposited at a much higher sedimentation rate, with much higher and more variable XRF core scanning Zr/Al ratios than the underlying sediment, and that show evidence of winnowing, particle sorting and increasing grain size, which we interpret to be related to the increasing flow of MOW. Pliocene sedimentary cyclicity is clearly visible in both the benthic $\delta^{18}\text{O}$ record and the Zr/Al data and is probably also precessionally controlled. Two contouritic bigradational sandy-beds are revealed above the third sedimentary cycle of the Pliocene. On the basis of these results, we conclude that sedimentation associated with weak Mediterranean–Atlantic exchange, began in the Gulf of Cadiz virtually at or shortly after the Miocene–Pliocene boundary.

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1. Introduction

Today, Mediterranean Outflow water (MOW) is the dominant intermediate water mass in the Gulf of Cadiz (Hernández-Molina et al., 2014a). The mixture of relatively warm and saline Mediterranean water and colder, less saline Eastern North Atlantic Central Water (ENACW) proceeds north and north-westwards along the middle slope of the Algarve Margin (Fig. 1; Hernández-Molina et al., 2003) and preconditions Atlantic Meridional Overturning Circulation, hence also influencing global climate (Ivanovic et al., 2014; Rogerson et al., 2012a).

The present day gateway configuration through the Gibraltar Strait dates back to the astronomically dated Miocene–Pliocene boundary (5.332 Ma; e.g. Lourens et al., 1996; Van Couvering et al., 2000; Blanc, 2002; Duggen et al., 2003; Roveri et al., 2014). Before this, Atlantic–Mediterranean seawater exchange took place through two marine connections, the Betic Corridor in southern Spain and the Rifian Corridor in north west Morocco (e.g. Benson et al., 1991; Santisteban and Taberner, 1983). Exchange became progressively restricted during the late Miocene as a result of a complex combination of tectonic and relative sea-level processes until the two water bodies became almost completely disconnected (Krijgsman et al., 1999; Kuroda et al., 2016; Lofi et al., 2005; Meijer and Krijgsman, 2005; Ohneiser et al., 2015; Simon and Meijer, 2015). Between 5.97 and 5.33 Ma, the chemical composition and salinity of Mediterranean water must have changed dramatically

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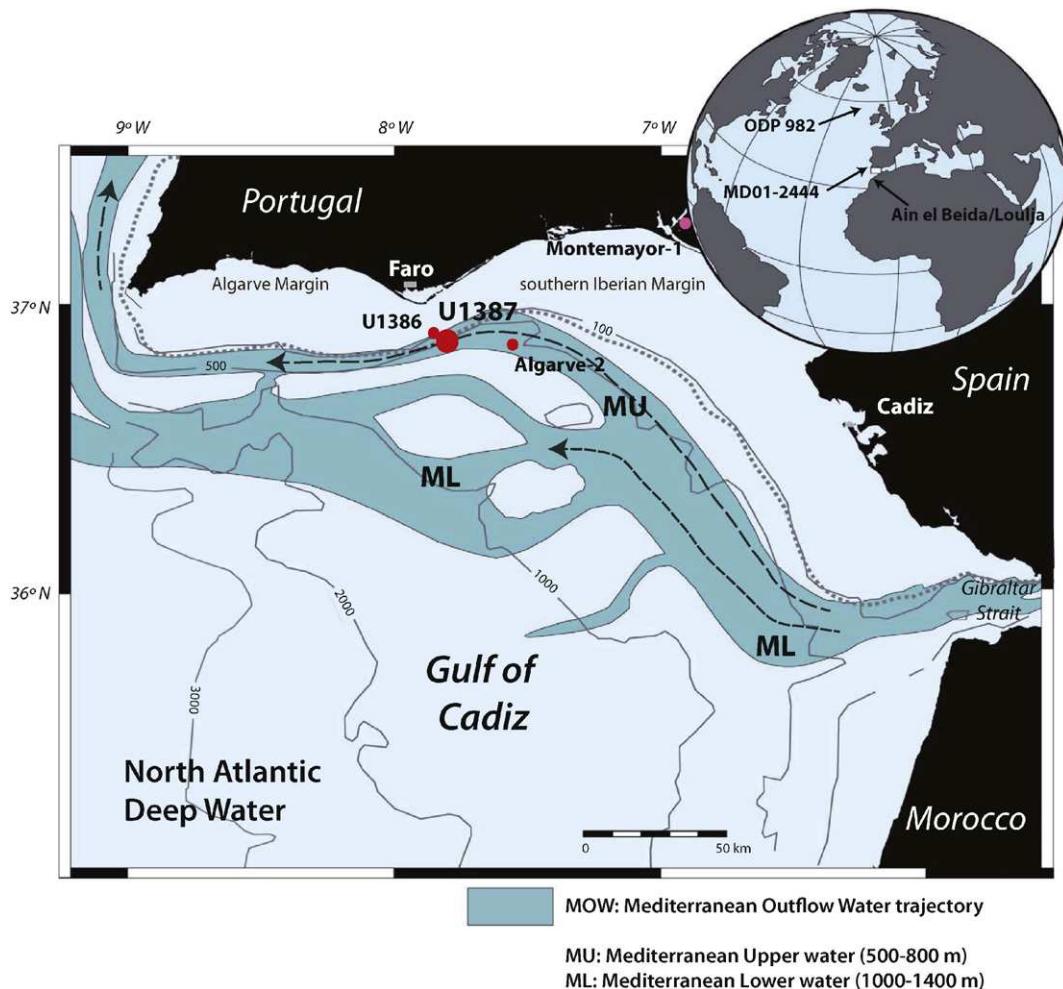


Fig. 1. Hydrography of the Gulf of Cadiz showing the main modern flow paths of MOW, specified as the Upper and Lower branch. Locations of IODP Sites U1386 and U1387, Algarve-2, Site ODP 982, MD01-2444, the onshore Montemayor-1 core in SW Spain and exposed sections of Ain el Beida/Loulja in NW Morocco are indicated. Modified after Hernández-Molina et al. (2003).

and thick evaporites precipitated in the Mediterranean Basin during an extraordinary event known as the Messinian Salinity Crisis (MSC; e.g. Hsu et al., 1973). Immediately after the Messinian, it is thought that physically the Mediterranean and the Atlantic were reconnected through the single gateway. However, the nature of the initial early Pliocene Atlantic–Mediterranean water mass exchange is still unclear (Hernández-Molina et al., 2014a).

Analyses of benthic foraminiferal assemblages indicate that water from the Atlantic Ocean penetrated the eastern Mediterranean within a few precession cycles above the Miocene–Pliocene boundary (Iaccarino et al., 1999a). This, combined with the transition back to normal marine salinities in the Mediterranean, suggests that exchange between the two water bodies took place through the Gibraltar Strait immediately after the Miocene–Pliocene boundary. Seismic reflection profiles, however, are interpreted to indicate that contourites associated with the earliest active bottom water current along the pathway of the MOW were deposited significantly later at 4.2–4.5 Ma (Expedition 339 Scientists, 2012, 2013a; Hernández-Molina et al., 2014b). The question remains whether any record of early Pliocene MOW prior to 4.2–4.5 Ma can be identified. Possible reasons for the absence of a clear early Pliocene record of MOW include the lack of appropriately located, well-dated sedimentary archives or that the MOW and ENACW are too similar to distinguish from each other (Rogerson et al., 2012b).

Integrated Ocean Drilling Program (IODP) Expedition 339 cored Hole U1387C (Fig. 1) in order to recover the first upper Miocene to lower Pliocene sediment record in the Gulf of Cadiz and to evaluate

the (re-)establishment of Mediterranean–Atlantic exchange after the MSC (Stow et al., 2011). Unfortunately, the Miocene–Pliocene boundary is not easy to identify in the Gulf of Cadiz, since its global stratotype has been established in the Mediterranean where it is associated with the facies shift from Messinian evaporites to Zanclean marls (Van Couvering et al., 2000). Consequently, the Miocene–Pliocene boundary outside the Mediterranean Basin can only be identified using alternative chronostratigraphic methods.

Analyses to detect bottom water currents, for example of Mediterranean–Atlantic exchange, are increasingly done by analysing X-ray fluorescence (XRF) scanning Zr/Al levels in combination with grain size analysis (Bahr et al., 2015; Kaboth et al., 2016; Lamy et al., 2015; Voelker et al., 2015b). In cases where Zr is associated with coarser grain sizes in marine environments, such as in the Gulf of Cadiz, it becomes a typical indicator for sediments affected by changes in bottom water current strength (Bahr et al., 2014; Bertrand et al., 2012; Ganeshram et al., 1999). The element Zr is most often associated with the dense refractory mineral zircon (e.g. McLennan et al., 1993). Zr/Al peaks are associated with higher proportions of non-aluminosilicate minerals, such as quartz, feldspars, and heavy minerals (zircon). Lower Zr/Al ratios are linked to finer grained Al-rich minerals, such as clays. Bottom water current sorting promotes the separation between coarser siliciclastic sediment components, enriched in Zr, and Al-rich finer grained minerals. For example, Bahr et al. (2014) showed that latest Pleistocene (0–140 ka) contourites at Site U1387 exhibited high Zr/Al ratios in sandy intervals, which were deposited under high bottom

water current flow conditions. By contrast, Zr/Al levels in the hemipelagic core MD01-2444 (Fig. 1) are dominated by long-term variability and shows only small-scale oscillations.

In this study, we establish a refined chronostratigraphic framework for upper Miocene to lower Pliocene sediments recovered at IODP Site U1387 based on seismic correlation, biostratigraphic constraints (planktic foraminifers, calcareous nannofossils), magnetostratigraphy and cyclostratigraphy. We then evaluate the imprint of bottom water currents on the Gulf of Cadiz sediments spanning the Miocene to Pliocene transition using lithology, grain size fractions ($>63 \mu\text{m}$) and elemental XRF scanning Zr/Al ratios. Our key aim is to identify deposits characteristic for elevated bottom water currents, to evaluate current strength, and if possible to date the onset of post-MSM MOW.

2. Background

IODP Site U1387 is located on the continental margin of the Iberian Peninsula at the eastern end of the Faro Drift ($36^{\circ}48'\text{N}$, $7^{\circ}43'\text{W}$) at 559 m water depth (Fig. 1; Expedition 339 Scientists, 2012, 2013a). This study focuses on the lowermost section of Hole U1387C from 731.20 mbsf to the base of the sediment sequence at 865.85 mbsf. These are the oldest sediments recovered during the IODP Expedition 339. Core recovery over this interval was 64% and downhole logging data was not collected because of borehole wall collapse, which means that the record contains significant gaps.

Using the last occurrence of *Globorotalia margaritae* (older than 3.85 Ma; Hilgen et al., 2012) at 560.4 mbsf, the first occurrence of *Globorotalia puncticulata* (older than 4.52 Ma; Lourens et al., 2004) at 630.8 mbsf in combination with the presence of *G. margaritae* at the base of the Hole (younger than 6.08 Ma; Hilgen et al., 2012) and the predominantly dextral coiling direction of *Neogloboquadrina acostaensis* (younger than 6.37 Ma; Hilgen et al., 2012; Table 1; see Section 5.1.1.), shipboard scientists suggested that sediments from 731.20 to 865.85 mbsf were Miocene/Pliocene in age (Expedition 339 Scientists, 2012; Stow et al., 2013; Hernández-Molina et al., 2014b). However, there is no robust justification of the depth for the Miocene–Pliocene boundary due to limited tie point identifications during the Expedition (Expedition 339 Scientists, 2012, 2013a). Although the exact depth of the boundary is never clearly stated in any of these publications, it is apparent from the various stratigraphic summary figures that so far the boundary was poorly constrained. The preferred depth of ~730 mbsf used in the summary in Hernández-Molina et al. (2014b), centred on IODP Expedition 339 results, was based on a lithologic transition from hemipelagites of presumed Miocene age to turbidites and debrites of presumed Pliocene age.

Recently, 3D and 2D seismic reflection profiles have been used to correlate the Algarve-2 well to IODP Hole U1387C (Fig. 1, Hernández-Molina et al., 2016). The Miocene–Pliocene boundary in Algarve-2 has been identified at 1455–1460 mbsf using a combination of biostratigraphic

data and cyclostratigraphic tuning of resistivity logs (Hernández-Molina et al., 2016). The Miocene–Pliocene boundary is manifested as an increase in reflection amplitudes on seismic lines (M-reflector) truncating towards the basin margins. This reflector was correlated in Hernández-Molina et al. (2016) to an abrupt sedimentary change in Site U1387 at ~826 mbsf with an uncertainty of 15 to 20 m.

Shipboard colour reflectance data (L^*) mirrors to some extent darker and lighter colours of core photographs (Fig. 2G). Core descriptions prepared during IODP Expedition 339 describe the lithology between 731.20 to 748.09 mbsf as greenish grey to very dark greenish grey nannofossil (silty) mud or sand with biogenic carbonate (Fig. 2A; Expedition 339 Scientists, 2013a). Cemented grey medium sandstone with biogenic carbonate is found between 748.10 and 750.92 mbsf (Fig. 2B). From 760 mbsf to the base of the core, dark greenish grey nannofossil mud and muddy/clayey nannofossil ooze is found commonly interbedded with 2 cm thick beds of dark greenish grey silty sand (Fig. 2D). The deepest occurrence of a thick (~80 cm) silty bed is observed at about 801 mbsf (Fig. 2C). Distinct bioturbation is present from 827.20 to 865.85 mbsf (Fig. 2E) and is less extensive higher up the sequence. Hence, compositional lithological changes exist at several depths in the studied section. While the Miocene–Pliocene boundary was previously presumed to be associated with the significant change in lithology at ~730 mbsf (Hernández-Molina et al., 2014b), it was just as likely represented by one of the other notable changes in lithology.

Finally, shallow water benthic foraminifers, such as *Ammonia*, *Elphidium* and *Asterigerinata* were recorded from 792.80 mbsf upwards and have particular high abundances above 765.72 mbsf (Expedition 339 Scientists, 2013a).

3. Methods

3.1. Sample preparation

Bulk sediment samples from IODP Hole U1387 of ~25 cm³ were collected at the IODP Bremen Core Repository for micropaleontological, stable isotope, and grain size analyses. Samples were taken at least every 40 cm from 731.20 to 7448.10 mbsf and every 20 cm from 751.12 to 865.85 mbsf. All samples were freeze-dried overnight, weighed, disaggregated in tap water overnight, sieved through >63 and $>150 \mu\text{m}$ sieves, dried, and weighed again. The $>63 \mu\text{m}$ fraction as a percentage of the total dry weight (wt%) was used as the sand fraction, that is in certain cases indicative for bottom water current flow strength (Rogerson et al., 2005).

Paleomagnetic samples consisted of the archive halves of all core sections and oriented discrete sediment samples, which were collected by inserting a hollow extruder into the middle of the working half of the split-core sections and then extruding the sediments into plastic cubes (2 cm \times 2 cm \times 2 cm, with an internal volume of ~7 cm³) as described in Expedition 339 Scientists (2013b).

Table 1

Planktic biostratigraphic events used in the present study and their position (mbsf) in IODP Hole U1387C and if it is stated in the preliminary IODP report (Expedition 339 Scientists, 2012). FO = First Occurrence, LO = Last Occurrence.

Planktic foraminifera biohorizon	Hole U1387C depth [mbsf]			Preliminary IODP report	Age bio-event [Ma]	Reference
	Top	Bottom	Mean			
LO <i>G. margaritae</i>	558.90	561.89	560.40	Yes	3.85	Lourens et al. (2004)
FO <i>G. puncticulata</i>	629.12	632.52	630.82	Yes	4.52	Lourens et al. (2004)
Influx <i>G. menardii</i>	752.40	751.30	751.85	No	5.55, 5.51, 5.31	Krijgsman et al. (2004), Van der laan et al. (2006)
Influx sinistral coiled <i>N. acostaensis</i>	808.22	809.52	808.87	No	5.30, 5.32	Lourens et al. (1996), Iaccarino et al. (1999b)
"Acme" <i>G. margaritae</i> > 10%	841.80 Not present (below base of the Hole)			No	5.70–5.85 –5.75–5.85	Van den Berg et al. (2015) Krijgsman et al. (2004)
Sinistral to dextral coiling change <i>N. acostaensis</i>	Not present (below base of the Hole)			Yes	6.37	Hilgen and Krijgsman (1999), Krijgsman et al. (2004), Sierra et al. (2001), Hilgen et al. (2012)

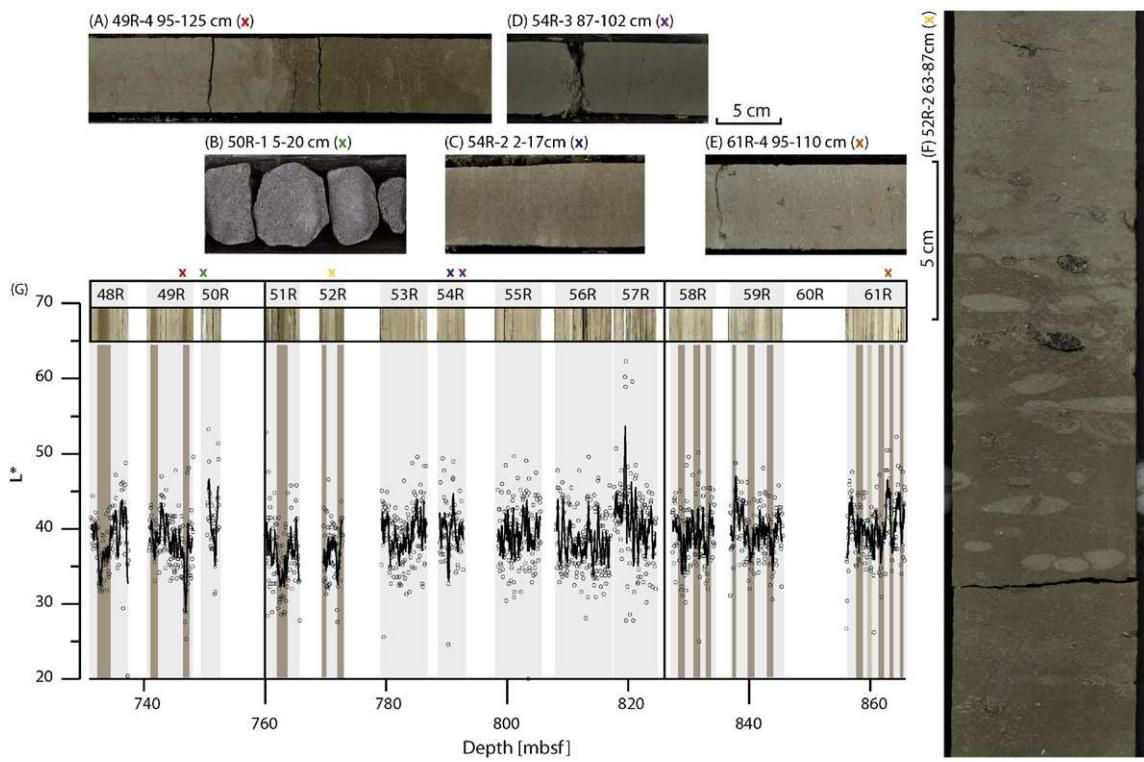


Fig. 2. Representative photographs of core sections from 731.20 to 865.85 mbsf (A–F). Coloured (x)s indicate their positions in the Hole, along with the IODP sample names that give the precise location of the image. (A) Sharp contact within greenish grey to very dark greenish grey nannofossil mud with colour change in a sandy interval. Bioturbation is visible (746.25 to 746.55 mbsf). (B) Cemented sandstone with biogenic carbonate (749.48 to 749.63 mbsf). (C) Homogenous sandier sediments (790.28 to 790.43 mbsf). (D) Thin 2 cm bed of dark greenish grey silty sand (792.61 to 792.76 mbsf). (E) Distinct bioturbation with smaller grain sizes (842.25 to 842.40 mbsf). (F) Small patches of sand preserved within a matrix of mud excavated by macrobenthic communities (771.62 to 771.86 mbsf). (G) 5-point moving average (black) of the colour reflectance (L^* ; open grey circles) is compared with darker colours of the core photographs with brown shades.

3.2. Micropaleontological studies

Planktic foraminiferal biostratigraphic analyses was performed on the $>150\text{ }\mu\text{m}$ fraction of the prepared samples. A microsplitter was used to split residues until about 150 planktic specimens remained. All planktic foraminiferal species were counted and classified including the biostratigraphic marker species *G. margaritae*, *Globorotalia menardii* and sinistral or dextral coiling *N. acostaensis*. Presence/absence data of benthic foraminiferal species on the $>150\text{ }\mu\text{m}$ fraction were used to provide a rough paleobathymetry estimation [a detailed quantitative analysis will be available in a separate publication by García-Gallardo et al., in preparation]. Five samples were chosen for detailed calcareous nannofossils biostratigraphic analysis, implementing the settling technique described in Flores and Sierro (1997).

3.3. Stable isotope analyses

Between two and fifteen specimens of benthic foraminifers *Cibicidoides pachyderma* $>250\text{ }\mu\text{m}$ were hand-picked from each sample for carbon and oxygen stable isotope analysis ($\delta^{13}\text{C}$, $\delta^{18}\text{O}$). Specimens were washed with ethanol, ultrasonicated for several seconds and dried for 24 h to remove all liquids.

Samples from 731.20 to 747.70 mbsf and 828.22 to 865.15 mbsf were analysed in the Godwin Laboratory for Palaeoclimate Research at the University of Cambridge. Isotope analyses were performed using a Thermo Finnigan MAT253 mass spectrometer fitted with a Kiel device. Analytical precision on an in-house standard, calibrated against international carbonate standard NBS-19, is $\pm 0.06\text{ }\text{\textperthousand}$ for $\delta^{13}\text{C}$ and $\pm 0.08\text{ }\text{\textperthousand}$ for $\delta^{18}\text{O}$. Samples from 751.12 to 824.34 mbsf were analysed in the Leibniz-

Laboratory for Radiometric Dating and Isotope Research at the Christian-Albrechts-University in Kiel with a MAT 253 mass spectrometer connected to a Kiel IV carbonate preparation device. The analytical precision of the NBS-19 international standard and three laboratory-internal standards was better than $\pm 0.05\text{ }\text{\textperthousand}$ for $\delta^{13}\text{C}$ and $\pm 0.08\text{ }\text{\textperthousand}$ for $\delta^{18}\text{O}$. At the Christian-Albrechts-University, large foraminiferal samples with more than six individuals were crushed to homogenize the samples before analysis of a representative subsample was undertaken. Results are reported relative to the Vienna Pee Dee Belemnite standard (VPDB).

3.4. X-ray fluorescence analyses

We used the XRF core scanner II (AVAAATECH Serial No. 2) at the MARUM—University of Bremen to retrieve elemental Zr/Al ratios. XRF core scanner data were collected every 3 cm down-core over a 1.2 cm^2 area with a down-core slit size of 10 mm in three separate runs using generator settings of 10, 30, and 50 kV, and currents of 0.2, 1.0, and 1.0 mA, respectively. Sampling time was set to 20 s and scanning took place directly at the split core surface of the archive half. The split core surface was covered with a $4\text{ }\mu\text{m}$ thin SPEX Certi Prep Ultralene1 foil to avoid contamination of the XRF measurement unit and desiccation of the sediment. The reported data have been acquired by a Canberra X-PIPS Silicon Drift Detector (SDD; Model SXD 15C-150-500) with 150 eV X-ray resolution and the Canberra Digital Spectrum Analyzer DAS 1000. The XRF scanner is equipped with an Oxford Instruments 50 W XTF5011. Core disturbance from drilling (biscuiting) may have significantly degraded the quality of XRF measurements from 760.18 to 764.94 (Core 51R) and inhibited measurements from 816.58 to 827.23 mbsf (Core 57R).

3.5. Magnetostratigraphy

As described in [Expedition 339 Scientists \(2013a, 2013b\)](#), continuous paleomagnetic measurements were made every 5 cm along the split-core sections during Expedition 339 using the shipboard cryogenic magnetometer (2G Enterprises Model-760R). The natural remanent magnetization (NRM) was measured after progressive alternating field (AF) demagnetization at 0, 10, and 20 mT. The split-core data were further cleaned by removing results from disturbed intervals and from near the ends of the core sections, as these data contain biases. The cleaned data are available in Table T16 of [Expedition 339 Scientists \(2013a\)](#) and the raw data are available online from the IODP LIMS Database.

To permit more detail demagnetization experiments and to better assess the characteristic remanent magnetization (ChRM) directions, we also collected approximately one discrete paleomagnetic cube sample (7 cm^3) from each core section (roughly one sample every 1.5 m) during Expedition 339 for the interval from 700 mbsf to the base of the hole, for a total of 56 samples. Only four of these samples were measured during Expedition 339. The remaining 52 were measured post-cruise in the Paleomagnetism Laboratory at University of California–Davis. All 56 samples were subjected to progressive AF demagnetization generally using 5 mT steps from 0 to 40 mT and 10 mT steps up to 100 mT. Paleomagnetic directions were determined from principal component analysis (PCA; [Kirschvink, 1980](#)) using ZPLOTIT software ([Acton, 2011, <http://paleomag.ucdavis.edu/software-zplotit.html>](#)). The remanence measurements from at least five demagnetization steps for each interval were fit to lines using only steps between 20 and 60 mT. PCA lines were fit using both the FREE option, in which the line is not required to pass through the origin of the plot, and the ANCHORED option, in which line is anchored to the origin ([Supplementary Tables S17–S19](#)). A Fisherian mean direction was also computed from the highest several demagnetization steps to estimate a stable end point (SEP). The paleomagnetic data and PCA results are provided in the Supplementary material.

For each sample, we determined a preferred inclination from the PCA FREE, PCA ANCHORED, or SEP estimate of the ChRM. The selection of which method best resolved the ChRM was based on visual inspection of the sample directions plotted on orthogonal vector diagrams (modified Zijderveld plots) and stereonets. The results for each sample were ranked with a quality factor, where Quality 1 was reserved for samples that displayed very linear decay of magnetization during demagnetization as noted by PCA maximum angular deviation (MAD) angles $<5^\circ$. Quality 2 are similar to Quality 1 but the MAD angles are between 5° and 10° . These highest quality (1 and 2) samples have very well resolved ChRM directions and the polarity can be determined with great confidence. Such high quality samples occur mainly in the upper part of section, and none are deeper than 550 mbsf. Quality 3 samples displayed moderate directional scatter during AF demagnetization but a relatively noisy ChRM can be determined from which the polarity can be established. Quality 4 samples have large directional scatter during AF demagnetization and generally a significant drilling overprint that overlaps much of the coercivity spectra of the ChRM. The ChRM cannot be accurately determined although it is often possible to determine the polarity of these samples because they trend towards positive (normal polarity) or negative (reverse polarity) inclinations as they are demagnetized above about 40–80 mT. Quality 5 samples have such highly variable directions during AF demagnetization or are so strongly overprinted that their polarity is indeterminate. Of the 56 samples below 700 mbsf, only 3 are Quality 3, 33 are Quality 4, and the remaining 20 are Quality 5.

4. Results

The sedimentary sequence was divided into three Intervals based on its lithological and geochemical properties ([Fig. 3](#)): Interval I (~826 to

865.85 mbsf) is characterized by low Zr/Al levels and consistently low wt.% sand ($>63 \mu\text{m}$); Interval II (~760 to ~826 mbsf) has higher, more variable Zr/Al ratios and progressively higher and more variable wt.% sand; Interval III (731.20 to ~760 mbsf) is characterized by highly variable Zr/Al oscillations and much higher sand grades commonly reaching 10–40 wt.%.

4.1. Sand fraction and composition

The sand fraction ($>63 \mu\text{m}$) of Interval I is consistently less than 3 wt.% ([Fig. 3](#)). Microscopic inspection of washed samples revealed that this coarser fraction mainly consists of planktic foraminifers with minor lithic particles including pyrite and rare quartz. A similar sand-grade and composition is observed in the lower part of Interval II (Cores 56R and 57R), but samples with distinctly higher wt.% of the $>63 \mu\text{m}$ fraction occur increasingly commonly above ~803 mbsf. The coarser grain size fraction is mainly associated with an increase in the quartz content. Sediments from 788.80 to 805.68 mbsf in Cores 54R and 55R show distinct bigradational patterns with traces of bioturbation ([Figs. 2D, 4](#)). The sand fraction of these silt horizons is dominated by fine sand-grade material (e.g. between 63 and 150 μm ; [Fig. 4](#)). The composition of the $>150 \mu\text{m}$ fraction is mainly restricted to foraminifers and peaks in the abundance of this coarser size fraction show no relationship with the bigradational pattern seen in the fine sand material ([Fig. 4](#)). The higher sand content seen in Interval III is associated with more lithic particles, mainly quartz grains ([Fig. 3](#)).

4.2. Stable isotope records

The benthic $\delta^{18}\text{O}$ isotope record shows regular fluctuations throughout the studied section. However, the different Intervals are characterized by variations with different wavelengths and amplitudes ([Fig. 3](#)). The $\delta^{18}\text{O}$ oscillations of Interval I have an average wavelength of $\sim 2.2 \text{ m}$, considering an additional fourth alternation in Core 61R that is only visible in the $\delta^{13}\text{C}$ during the larger $\delta^{18}\text{O}$ oscillation from 856.68 to 860.48 mbsf ([Table 2; Figs. 3, 9](#); see [Section 5.2.2.](#)) [a detailed interpretation of the $\delta^{13}\text{C}$ record is beyond the scope of this paper and is the subject of a forthcoming paper ([Van der Schee et al. in preparation](#))]. The regular $\delta^{18}\text{O}$ alternations of Interval II typically have significantly longer wavelengths of $\sim 5.9 \text{ m}$ on average. Limited recovery of Interval III makes the wavelength of its $\delta^{18}\text{O}$ alternations more difficult to calculate, but it appears to be comparable to those of Interval II. The average $\delta^{18}\text{O}$ of Interval II is lower than that of Interval I and the average of Interval III is lower than Intervals I and II. In all three Intervals, distinctly darker layers tend to be associated with higher $\delta^{18}\text{O}$ ([Fig. 3](#)).

4.3. XRF scanning Zr/Al ratios

Zr/Al ratios of Interval I have consistently low ratios with the lack of obvious regular alternations that resemble the periodicity of those well developed in the $\delta^{18}\text{O}$ ([Fig. 3A, B](#)). By contrast, Intervals II and III are characterized by distinct, high amplitude variability in Zr/Al levels displaying oscillations with wavelengths similar to those of the benthic isotope record ($\sim 5.9 \text{ m}$). In general in Interval II and III, these two records show negative co-variation.

4.4. Paleobathymetry estimations from benthic foraminiferal presence/absence data

Presence/absence data of benthic foraminifers are provided in the Supplementary material. The studied samples frequently contain benthic foraminifers known to have their upper limit of bathymetric distribution at upper to middle bathyal water depths (e.g., *C. pachyderma*, *Cibicidoides wuellerstorfi*, *Laticarinina pauperata*, *Eggerella bradyi*,

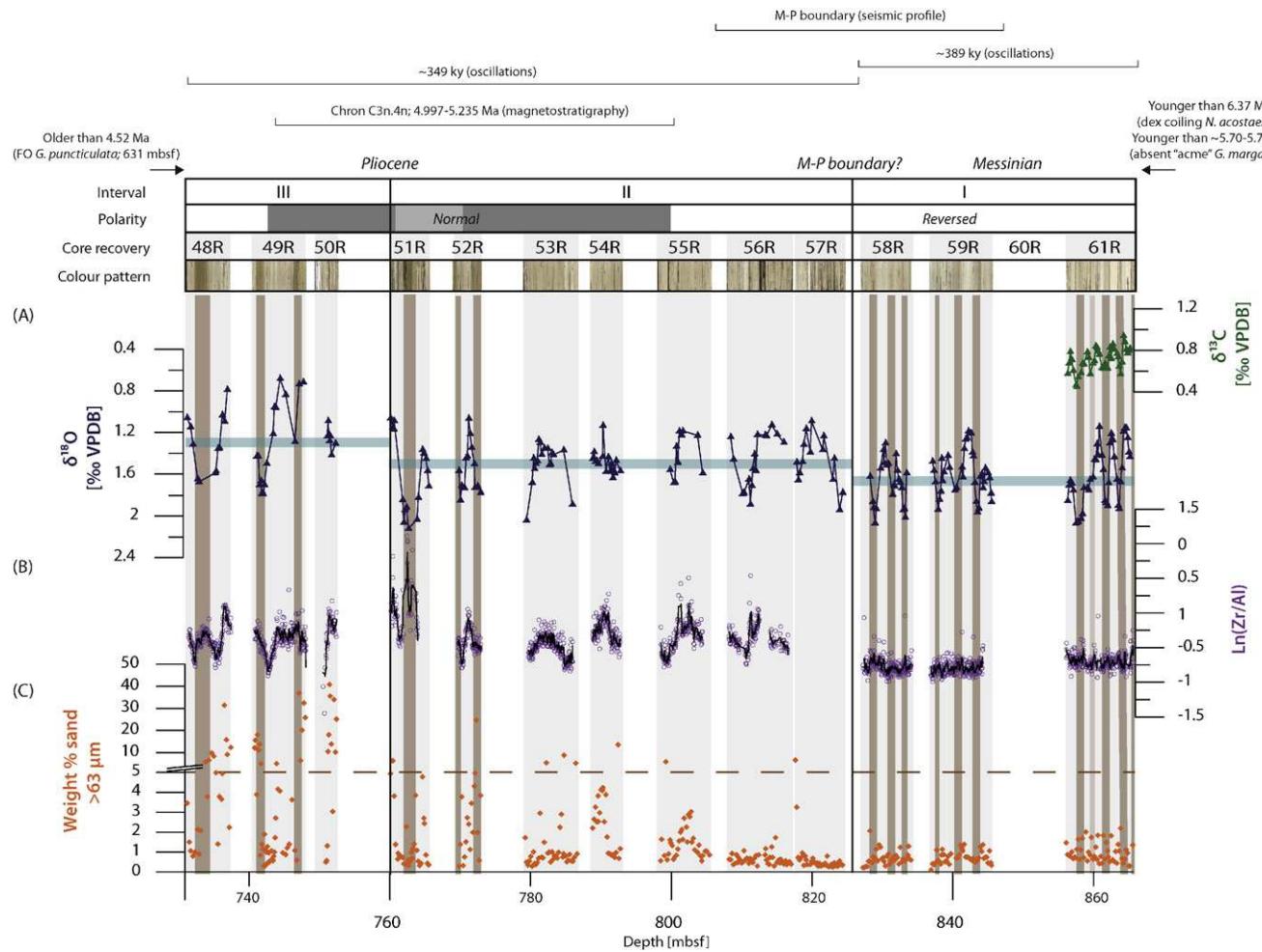


Fig. 3. Paleoclimate records of IODP Site U1387C. Intervals I, II and III, magnetostratigraphic polarities, core numbers and their recovery, and colour pattern from core photographs are shown above the charts (Expedition 339 Scientists, 2013a). Brown shaded areas indicate darker colours in core photographs. Age indications are displayed above the panel. FO = First Occurrence. (A) $\delta^{18}\text{O}$ (blue) and $\delta^{13}\text{C}$ (green) of benthic *C. pachyderma*. Light blue horizontal bars show average $\delta^{18}\text{O}$ values per Interval. (B) Elemental Zr/Al ratios (open purple circles) with 5-point moving average (black). (C) Weight % sand $>63\text{ }\mu\text{m}$ as a percentage of total dry weight (orange).

Epistominella exigua, *Fissurina* spp., *Gyroidinoides* spp., *Melonis* spp., *Oolina* spp., *Osangularia culter*, *Pullenia* spp., *Saracenaria* spp.; Jones, 1994; Leckie and Olson, 2003; Van Morkhoven et al., 1986). Furthermore, the presence/absence data show taxa that characterize slope environments between 300 and 900 m water depth in the Gulf of Cadiz and the western Iberian Margin (e.g., *Amphicoryna scalaris*, *Bigenerina nodosaria*, *Bulimina striata*, *Cibicidoides mundulus*, *Globocassidulina subglobosa*, *Planulina ariminensis*, *Uvigerina mediterranea*, and *Uvigerina pygmaea*) (Phipps et al., 2012; Schönfeld, 1997, 2002). A particularly interesting species among the bathyal taxa is *L. pauperata*, a cosmopolitan species that tolerates a wide range of environmental conditions at bathyal and abyssal water depths (Van Morkhoven et al., 1986). While other taxa limited to bathyal water depths are present in Interval I as well as Interval II and III, this species disappears from the record above ~ 826 mbsf. An upper depth limit at the middle to upper bathyal transition has been suggested for *L. pauperata*, and its disappearance may indicate a slight shallowing from middle to upper bathyal water depths comparable to today (Van Morkhoven et al., 1986).

Foraminiferal species restricted to neritic environments such as *Ammonia* spp., *Asterigerinata* spp., *Cancris* spp., and *Elphidium* spp. are increasingly present above ~ 801 mbsf (Leckie and Olson, 2003; Mendes et al., 2012). These taxa occur throughout the record alongside bathyal species mentioned above and are thus considered allochthonous.

5. Chronostratigraphic framework

5.1. Bio- and magnetostratigraphy

5.1.1. Planktic foraminifers

Several planktic foraminiferal species provide constraints on the age of the basal part of IODP Hole U1387 (Table 1). These include the following:

1. The first occurrence of *G. puncticulata* (4.52 Ma; Lourens et al., 2004) which has been identified in IODP Hole U1387C at a depth of 630.82 mbsf (Expedition 339 Scientists, 2013a). The interval of interest (731.20 to 865.85 mbsf) is therefore entirely older than 4.52 Ma.
2. *N. acostaensis* which is abundant in upper Messinian to lower Pliocene sediments from this region and exhibits well-dated changes in predominant coiling direction and short time scale influxes of the inferior coiling direction (Krijgsman et al., 2004; Sierro et al., 2001; Lourens et al., 2004). Typically, assemblages strongly dominated by sinistral coiling *N. acostaensis* are found in sediments older than 6.37 Ma, while dextral coiling shells constitute the vast majority of specimens in sediments younger than 5.82 Ma (Hilgen and Krijgsman, 1999; Krijgsman et al., 2004; Sierro et al., 2001). Quantitative assessment of the coiling direction of *N. acostaensis* in the lower part of Hole U1387C shows that it is strongly dominated by dextral coiling forms (Fig. 5). Consequently, the base of the Hole is most likely younger than 5.82 Ma.

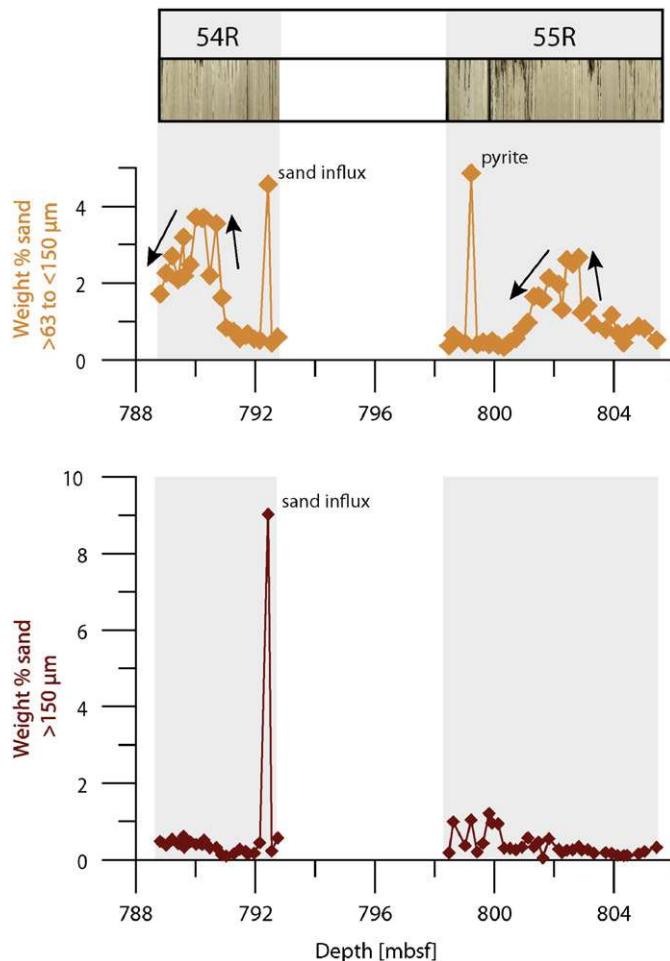


Fig. 4. Weight % sand of the fine sand-grade (>63 to <150 μm ; orange) and coarser sand-grade (>150 μm ; red) fractions from 788.80 to 805.68 mbsf. Core numbers and photographs are displayed above the graphs. Two outliers are explained by the presence of pyrite and a centimetre scale sand influx (visible on core photographs; Fig. 2D). The distinct increase and decrease in the fine sand-grade fraction are indicating the bimodal pattern (accentuated with black arrows).

Two sinistral coiled influxes of *N. acostaensis* are found from 808.22 to 809.52 and at 841.80 mbsf (Fig. 5). In the Mediterranean Basin in the lower Pliocene, two sinistral influxes are reported and astronomically tuned in the Eraclea Minoa section to 5.30 and 5.32 Ma (Lourens et al., 1996). Equivalent influxes to these are described in precessional driven Cycles 2 and 3 immediately above the Miocene–Pliocene boundary in ODP Holes 974B and 975B (Iaccarino et al., 1999b).

Table 2
Wavelengths of cyclical alternations of $\delta^{18}\text{O}$ for selected intervals.

Core	Hole U1387C depth [mbsf]		Length [m]	Cycles [-]	Length/cycle [m/cycle]
	Top	Bottom			
48R	731.20	736.91	5.71	<1	–
54R-57R	790.44	819.91	29.47	5.0	5.89
58R	828.92	832.92	4.00	2.0	2.00
59R	837.90	844.50	6.60	2.5	2.64
61R	856.68	864.42	7.74	3.0 + 1 ^a	1.94
		Weighted average		2.16	
Interval I (58R-61R):					

^a Three cyclical alternations are visible in the $\delta^{18}\text{O}$ record plus an additional cycle visible in $\delta^{13}\text{C}$ is taken into consideration.

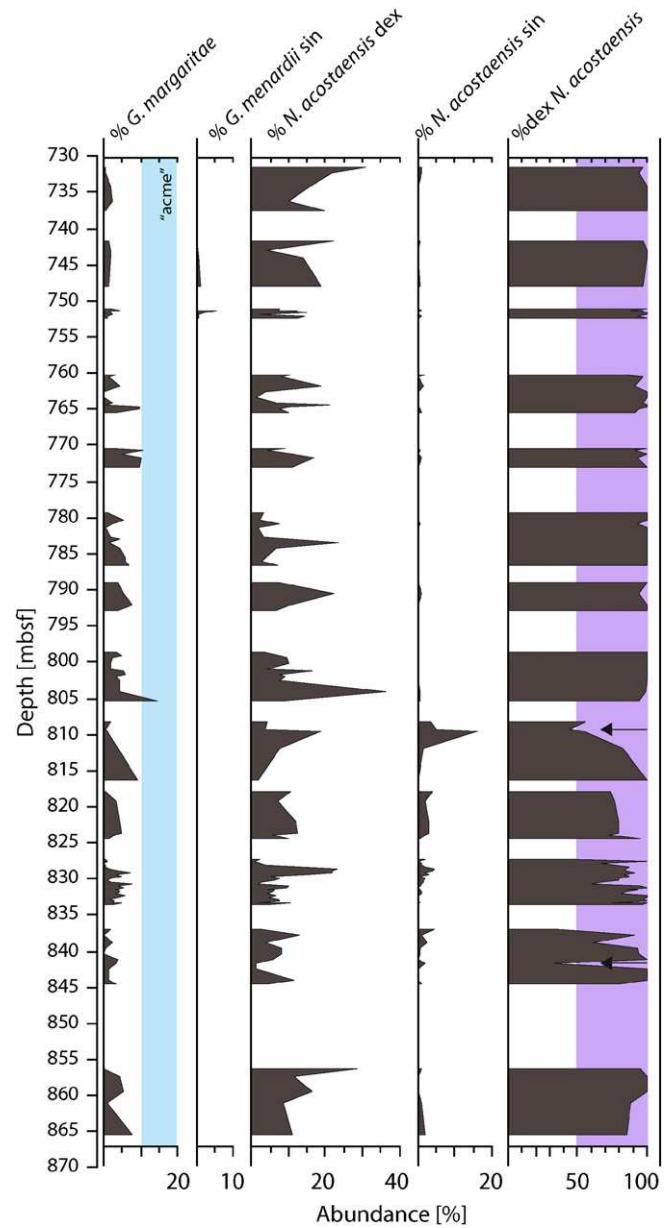


Fig. 5. Planktic foraminiferal biostratigraphy of Hole U1387C. Relative abundances are displayed. The light blue shaded area shows the “acme”-zone of *G. margaritae*. The %dex *N. acostaensis* displays the relative distribution of dex/(dex + sin) of *N. acostaensis*. The purple shaded area indicates that most species have dextral coiling. The two arrows highlight the sinistral coiled influxes.

3. A distinct high abundance interval, referred to as “acme”, of *G. margaritae* has been described from several upper Miocene records from the Gulf of Cadiz, SW Spain and NW Morocco (Van den Berg et al., 2015; Krijgsman et al., 2004; Ledesma, 2000; Sierro et al., 1982, 1993). The “acme” has a relative abundance of *G. margaritae* that exceeds at least 10% of the total planktic foraminiferal assemblage. An interval of particular high abundance occurs in the Ain el Beida section (>20%; NW Morocco; Krijgsman et al., 2004) and the Montemayor-1 core (>37%; SW Spain; Van den Berg et al., 2015) between respectively ~5.75 and ~5.70 Ma until ~5.84 Ma (both sites), where the “acme” coincides with glacial stages TG20 and TG22. Foraminiferal counts from the lower part of Hole U1387C show that only one sample contains >10% *G. margaritae* (805.42 mbsf; Fig. 5), indicating that this sedimentary succession

does not overlap the high abundance “acme”. The relative large time span of the “acme” seems to be too long to coincide within a coring gap (see Section 5.2.2.). In addition, it is unlikely that the “acme” would be absent even though the sediments of this time interval might have been covered, since it has always been found in Gulf of Cadiz successions. Since the coiling direction of *N. acostaensis* indicates that the studied sediments are most likely younger than 5.82 Ma, the *G. margaritae* data further constrains the age of the sediments to being younger than ~5.70–5.75 Ma. Outside the very high abundance interval, Krijgsman et al. (2004) included all samples with >5% *G. margaritae* in the wider “acme” interval that spans 5.56 to 5.84 Ma. As our quantitative *G. margaritae* data typically varies between 2 and 10% (Fig. 5; Supplementary material), it is possible that these sediments at the base of IODP Hole U1387C overlap, at least in part, with the described 5.56 to 5.84 Ma interval.

4. An influx of *G. menardii* with a maximum relative abundance of 5.5% is recorded between 751.33 and 752.28 mbsf (Fig. 5; Supplementary material). Three similar influxes have been documented from the latest Messinian to earliest Pliocene in the Ain el Beida and Loulja sections in NW Morocco with astronomical ages of 5.55, 5.51 and 5.32 Ma.¹ Without more robust age constraints the influx of *G. menardii* in IODP Site U1387 cannot be correlated to any of these.

In summary, planktic biostratigraphic results suggest that the interval from 731.2 to 865.85 mbsf in IODP Hole U1387C is older than 4.52 Ma, the first occurrence of *G. puncticulata* (630.82 mbsf) and younger than the high abundance *G. margaritae* “acme” at ~5.70–5.75 Ma.

5.1.2. Calcareous nannofossils

Characteristic calcareous nannofossil marker species from the latest Neogene are common with individuals of *Amaurolithus primus*, *Discoaster surculus*, *Discoaster pentaradiatus* and *Discoaster quinqueramus* as well as characteristic reticulofenestrids (*Reticulofenestra pseudoumbilicus*, *Reticulofenestra minuta*, *Reticulofenestra minutula* and *Reticulofenestra rotaria*), allowing to place the studied interval between biozones NN11 and NN15 spanning from 3.81 to 7.36 Ma (e.g. Young et al., 1994; Raffi et al., 2006). The presence of reworked specimens (e.g. *Cyclicargolithus floridanus*, *Discoaster deflandei*, *Sphenolithus belemnos* gr.), mainly of Paleogene and lower to middle Miocene varieties, is common (Raffi et al., 2006). Expedition 339 Scientists (2013a) report the last occurrence of *D. quinqueramus* at 811.43 mbsf what indicates an age younger than 5.54 Ma (Raffi et al., 2006). However, this biostratigraphic event is hard to identify in such detail, because only a few specimens of *D. quinqueramus* are present and thus could easily be reworked. Also, a few specimens of *R. rotaria* are found in the studied interval which indicate an age range between 6.79 and ~7.41 Ma (Flores et al., 2005; Lourens et al., 2004; Young et al., 1994). The planktic foraminiferal bio-events are clearly indicating upper Messinian and not upper Tortonian to lower Messinian as indicated by *R. rotaria*. If *R. rotaria* and planktic foraminifers were in-situ, specimens of *G. miotumida* and continuous abundances of *G. menardii* 5 were expected (Lourens et al., 2004). The incompatibility of the nannofossil and foraminiferal data suggest that the few specimens of *D. quinqueramus* and *R. rotaria* are reworked.

5.1.3. Magnetostратigraphy

The paleomagnetic data is generally of low quality because the sediment magnetizations are relatively weak and the magnetization of the cores includes a significant drilling overprint that masks the original depositional magnetization. The overprint has a steep downward direction (positive inclinations) similar to that expected for normal polarity intervals. Hence, failure to remove the drilling overprint with magnetic

cleaning, such as AF demagnetization, can bias the magnetic polarity interpretation significantly. This is particularly the case for the split-core sections, which give virtually only positive inclinations even in intervals that must be of reversed polarity. The 20-mT peak-field AF demagnetization used during Expedition 339 was clearly insufficient in removing the drilling overprint and resolving the original depositional magnetization, and thus no polarity could be determined from this data below 500 mbsf.

The discrete samples were subjected to much higher peak-field AF demagnetization, which was successful at removing the drilling overprint from many of the samples from Hole U1387C. Even though the quality of the samples is generally considered poor for resolving the ChRM fully, the polarity of the samples could be reasonably well estimated. This is illustrated in Fig. 6 where the inclinations are shown for the interval from 550 mbsf to the base of the Hole, along with the interpretation of the magnetozones.

Most importantly for the present study is that the stratigraphic section below 800.65 mbsf is all reversely magnetized. Based on the biostratigraphic constraints, this magnetozone corresponds to Chron C3r (5.235–6.033 Ma; Lourens et al., 2004). The base of this magnetozone was not recovered, which is consistent with the biostratigraphic constraints that suggest the age of the bottom of the hole is <5.75 Ma. The overlying normal polarity magnetozone spanning 743.99 to 800.65 mbsf corresponds to Chron C3n.4n (4.997–5.235 Ma; Lourens et al., 2004), indicating the boundary between the magnetozones at 800.65 mbsf has an age of 5.235 Ma. The Miocene–Pliocene boundary is thus below 800.65 mbsf and above the base of the Hole (865.85 mbsf). The sequence of magnetozones above Magnetozone C3n.4n fits well the general chron sequence of the geomagnetic polarity timescale and the depths and dates of the observed reversals agree well with biostratigraphic constraints and give sedimentation rates compatible with the other chronostratigraphic constraints (Fig. 7).

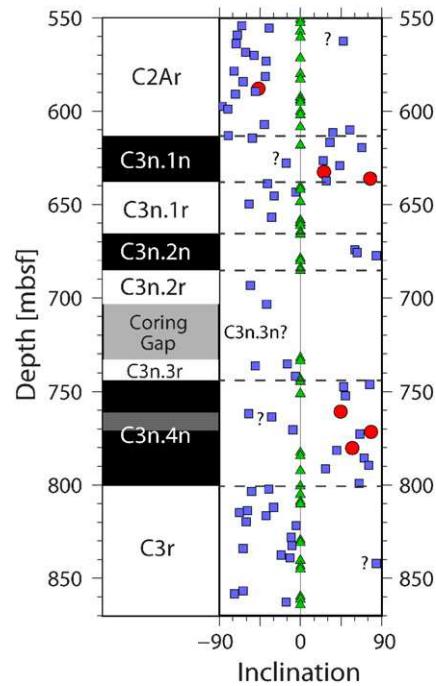


Fig. 6. Magnetostatigraphic interpretation for magnetozones between 550 mbsf and the base of Hole U1387C. The inclinations plotted are the preferred inclinations determined by principal component analysis (PCA), and are given in the Supplementary material. The symbols reflect the different quality of the results as discussed in the text and Supplementary material Quality 3 (red circles), Quality 4 (blue squares), and Quality 5 (green triangles), which are plotted along the 0° inclination merely to show the depth at which they occur.

¹ Please note that Krijgsman et al. (2004) incorrectly identify *G. menardii* pulses as *Globorotalia miotumida*. (Tulbere et al., in preparation), while they are referred to as *G. menardii* in van der Laan et al. (2006) (Van den Berg et al., 2015).

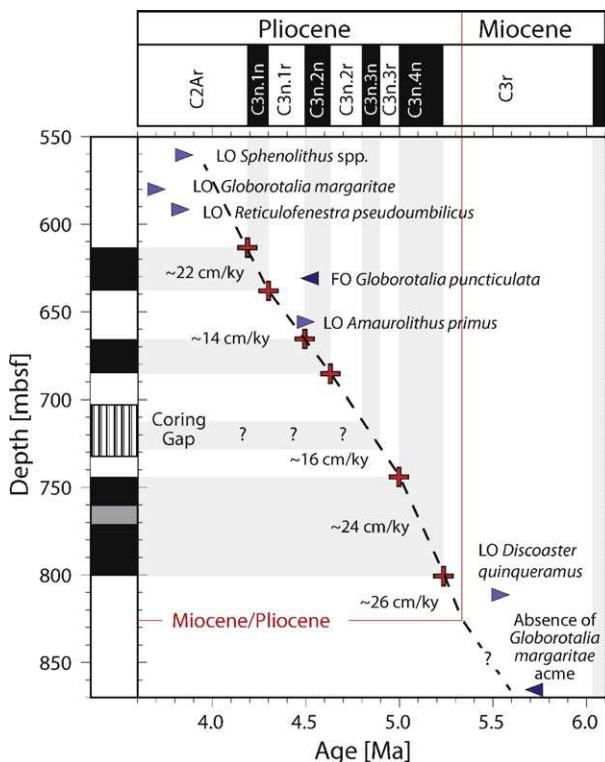


Fig. 7. Chronostratigraphy for Site U1387 from magnetostratigraphic and biostratigraphic constraints. FO = First Occurrence, LO = Last Occurrence.

5.2. Cyclostratigraphic constraints

5.2.1. Precessionally forced variations

Benthic $\delta^{18}\text{O}$ records in open ocean deep marine settings during the Pliocene–Pleistocene are mainly dominated by obliquity. However, the first ~150 ky at the base of the Pliocene show a strong precessional signal in the LR04 stack (Fig. 8B; Lisicki and Raymo, 2005). In contrast, rhythmic sedimentary cycles, including resistivity, gamma ray and Ti/Al records from the Iberian and Moroccan Atlantic margins during the Messinian and early Pliocene typically display a pronounced variability in response to precession (e.g. Hodell et al., 1994; Krijgsman et al., 2004; Pérez-Asensio et al., 2013, 2014; van der Laan et al., 2005, 2006, 2012; van den Berg et al., 2015). Precession continues to be dominant as astronomical forcing in IODP Sites drilled in the Gulf of Cadiz throughout the Pleistocene, including Site U1387 (Bahr et al., 2015; Lofi et al., 2016; Voelker et al., 2015b).

Precession and obliquity patterns are further investigated by comparing the $\delta^{18}\text{O}$ of the Loulja section with obliquity and a combined record of normalized precession minus normalized tilt (obliquity), hereafter precession-tilt (Fig. 8A, B). Obliquity is expressed in precession-tilt by the enhancement of the amplitude of every other precession cycle. This distinct pattern is clearly visible in the Loulja $\delta^{18}\text{O}$ record, for example around interglacial stage TG9 (5.45 Ma; Van der Laan et al., 2006). Here, the expression of TG9 is more pronounced due to increased influence of obliquity, while the precession cycles before and after TG9 have lower amplitudes.

The discontinuous nature of the record in the studied interval of Hole U1387C makes it impossible to assess the oscillation patterns in the same way as can be achieved for continuous successions. However, in Interval I, a few continuous regular alternations are visible mainly in the $\delta^{18}\text{O}$. Benthic $\delta^{18}\text{O}$ oscillation patterns in Cores 61R, 59R and 58R, show alternations of high and low amplitudes (Fig. 9), that are very similar to the typical interference patterns of obliquity and precession. To illustrate this, we show a comparison of the benthic $\delta^{18}\text{O}$ with a theoretical cyclical precession-tilt curve. A similar, close relationship

between the $\delta^{18}\text{O}$ benthic record and the precession-tilt curve is seen at the Loulja section (Fig. 8A, B).

In even more detail, Core 61R reveals three full regular alternations in the $\delta^{18}\text{O}$ record, while the $\delta^{13}\text{C}$ from the same foraminiferal specimens shows four regular alternations (Figs. 3, 9). Normally, ventilation patterns and organic carbon storage explain $\delta^{13}\text{C}$ variations. In the Gulf of Cadiz $\delta^{13}\text{C}$ variations have been explained as an expression of ventilation patterns linked to MOW during the Pleistocene with a strong precessional component (Voelker et al., 2006, 2015b), however, linking our data with these patterns without a definite understanding of the Atlantic–Mediterranean exchange is problematic. Independently, other well-dated benthic Messinian to upper Pliocene benthic- $\delta^{13}\text{C}$ records in the area possess a precession component, i.e. Ain el Beida and Loulja sections (Van der Laan et al., 2006, 2012). Consequently, it is likely that the four $\delta^{13}\text{C}$ regular alternations represent precession, while the long $\delta^{18}\text{O}$ alternation from 856.68 to 860.80 is the expression of diminished precession under a strong influence of obliquity. This suggests that the short scale visible regular alternations of the $\delta^{18}\text{O}$ record within Interval I are precession-induced cycles, with influence of other astronomical cyclicities.

Regular alternations have larger wavelengths in Interval II and III, while coring gaps are still present. As a result, comparing cycle patterns as done for Interval I is impossible, because only one or two cycles can be recognized between gaps. We assume that the $\delta^{18}\text{O}$ cyclic patterns are representing the same astronomical variations as in Interval I. Consequently, precessionally forced variability is assumed for the colour changes, Zr/Al ratios and, to some extent, grain size fraction analyses that have wavelengths consistent with the benthic $\delta^{18}\text{O}$ record.

5.2.2. Depositional time span

If the cycles in IODP Hole U1387C are precessionally controlled, the depositional time span can be calculated given that the average duration of a precession cycle is ~21.7 ky (Berger, 1984) and using the same sedimentation rate for intervals where no core was recovered as is measured in the cores below and above. No significant stretching or shortening of the sediment inside the core pipe during drilling is anticipated because the sediment throughout the Hole is well compacted.

In Interval I, Cores 58R, 59R and 61R all show distinct colour and $\delta^{18}\text{O}$ cycles with an average wavelength of 2.2 m per cycle (Table 2). Assuming this constant sedimentation rate, Interval I (38.68 m) should equate to ~18 precession cycles, with an average sedimentation rate of about 10 cm/ky and a time span of ~389 ky. The occurrence of two colour cycles and two carbon isotope excursions from 856.9 to 860.3 suggest the presence of two precession cycles during the long $\delta^{18}\text{O}$ cycle (Figs. 3, 9).

In Interval II, cores with more continuous recovery (Cores 57R to 54R) show an average thickness of 5.89 m per cycle (Table 2). Interval III appears too short to calculate an individual sedimentation rate in the same way, but it is clear that Core 48R has one long $\delta^{18}\text{O}$ -variation with a width of 5.71 m, which is presumably representing nearly a full cycle. Uncertainty increases due to a possible additional, subtly expressed cycle in Core 48R, as is hinted in the Zr/Al oscillations. However, this still suggests that the sedimentation rate throughout Interval II and III remains relatively constant. Thus, the total thickness of Interval II and III (97.72 m) equates to ~16 precession cycles, with an average sedimentation rate of about 27 cm/ky and a total depositional time span of ~349 ky.

5.3. Tentative working hypothesis on the age model for IODP Hole U1387C

While individually, none of these different datasets provide robust age constraints, more confidence can be gained if the stratigraphic interpretation is compatible with all of them. Here, we explore the best possible stratigraphic fit between the data available.

Assuming, on the basis of the seismic correlation (Hernández-Molina et al., 2016), that the Miocene–Pliocene boundary occurs at the abrupt lithological change between Interval I and II (826 mbsf),

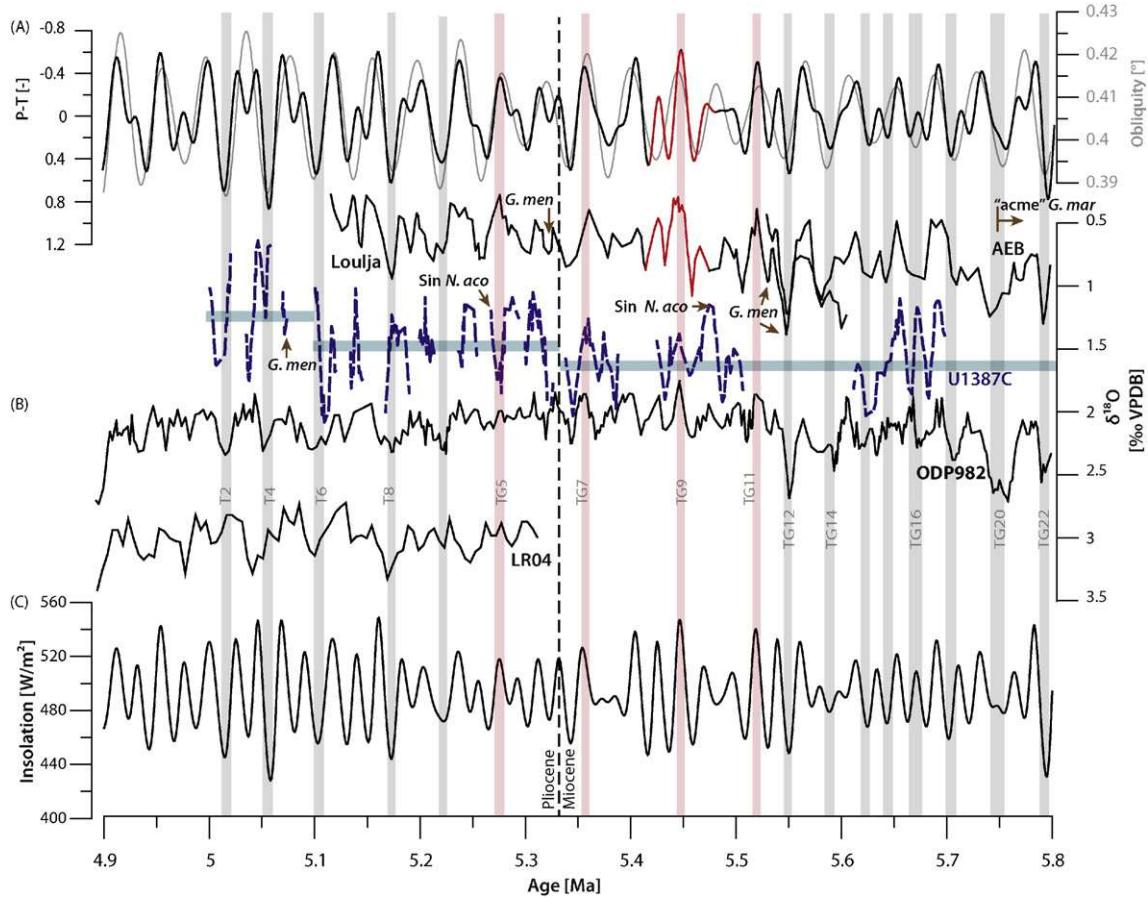


Fig. 8. (A) Normalized precession-tilt (obliquity; P-T; black) and obliquity (grey) of solution La04 (65°N summer insolation; Laskar et al., 2004), red highlights the clear resemblance with the Loulja $\delta^{18}\text{O}$ record. (B) Benthic $\delta^{18}\text{O}$ record of *Cibicidoides pachyderma* of Hole U1387C (dotted blue) for the constructed tentative age model with its three tie-points at 5.0, 5.33 and 5.7 Ma. Horizontal boxes show the average of IODP U1387 for the different Intervals. The $\delta^{18}\text{O}$ records of *P. ariminensis* of Ain el Beida (Krijgsman et al., 2004; Van der Laan et al., 2005), Loulja (Van der Laan et al., 2006) and ODP982 (corrected according to Van der Laan et al., 2006; Hodell et al., 2001) and the LR04-stack (Lisiecki and Raymo, 2005) are displayed as references. Typically, there is a strong 1:1 correlation between *P. ariminensis* and *C. pachyderma*, especially for $\delta^{18}\text{O}$ (e.g. Van der Laan et al., 2006; Voelker et al., 2015b). Influxes of *G. menardii* (G. men) and the "acme" of *G. margaritae* (G. mar) in Loulja, Ain el Beida and Hole U1387C are indicated with arrows as well as the sinistral coiled influxes of *N. acostaensis* (sin N. aco). TG stages are taken from Site ODP 846 (Shackleton et al., 1995), while T stages are used from the LR04-stack. (C) Insolation curve of solution La04 (65°N summer insolation; Laskar et al., 2004).

the duration of the ~18 precession cycles (389 ky) in Interval I suggests that the age of the base of Hole U1387C is ~5.72 Ma (age indications are above the panel in Fig. 3). This age is consistent with the absence of the *G. margaritae* "acme" which suggests that the bottom of the Hole must be younger than ~5.70–5.75 Ma. The minimum time span of the "acme" of at least 5 precession cycles makes it highly unlikely that the "acme" falls exactly in a coring gap. Within these age estimations, the two documented influxes of *G. menardii* are likely to be in the coring gap, within unrecovered Core 60R (5.55 and 5.51 Ma; stratigraphic planktic foraminiferal influxes and coiling changes are represented in Fig. 8B).

The depositional time span calculated on the basis of precessional cycles of Interval II and III (16 precession cycles; ~349 ky) suggests that, if the Miocene–Pliocene boundary is positioned right above Interval I, the age of the top of the studied interval is ~4.98 Ma. The precession-constrained sedimentation rates can also be used to estimate the age of the normal polarity interval from 743.99 to 800.65 mbsf. This suggests an age range for the normal polarity interval from ~5.04 to ~5.24 Ma. This is in good agreement with the Chron C3n.4n (4.9975.235 Ma; Lourens et al., 2004). The sinistral coiled influx of *N. acostaensis* from 808.22 to 809.52 mbsf can be closely linked to the second reported dextral to sinistral coiling change reported at 5.30 Ma in the Mediterranean Basin, three precession cycles above the Miocene–Pliocene boundary (Iaccarino et al., 1999b; Lourens et al., 1996). The influx of *G. menardii* between 752.4 and 751.3 mbsf, with its age estimation from this study of about 5.07 Ma, is thus younger than the influxes recorded in NW

Morocco sections (5.32 Ma; Van der Laan et al., 2006; Fig. 8B). Planktic foraminiferal biostratigraphy of the equivalent section in Morocco (Loulja) has not been studied in detail and therefore the coiling changes in *N. acostaensis* and influx of *G. menardii* may have been overlooked so far. The well-recorded influx of *G. menardii* at 5.32 Ma in the Loulja section (Van der Laan et al., 2006) is most likely lost in the coring gap just after the Miocene–Pliocene boundary (Fig. 8B).

In summary, the stratigraphic interpretation which places the Miocene–Pliocene boundary at the sharp sedimentological change at about 826 mbsf, in accord with the seismic correlation (Hernández-Molina et al., 2016), and using the precession-forced, sedimentary cycles to constrain the age above and below that point, provides an age model for the studied interval of Site IODP U1387. This age model is compatible with the available biostratigraphic and paleomagnetic data, which results in an age model with three tie-points: (1) an age of about 5.7 Ma at the base of the Hole, (2) the Miocene–Pliocene boundary (5.32 Ma) at about 826 mbsf and (3) an age of about 5.0 Ma for the top of the studied interval at 731.20 mbsf. An uncertainty of two to three precessional cycles should be anticipated for the base and the top of the studied interval given the methods used to construct the age model.

6. Discussion

Hemipelagic nannofossil muds showing significant bioturbation and containing few lithic coarser particles characterize the Messinian part

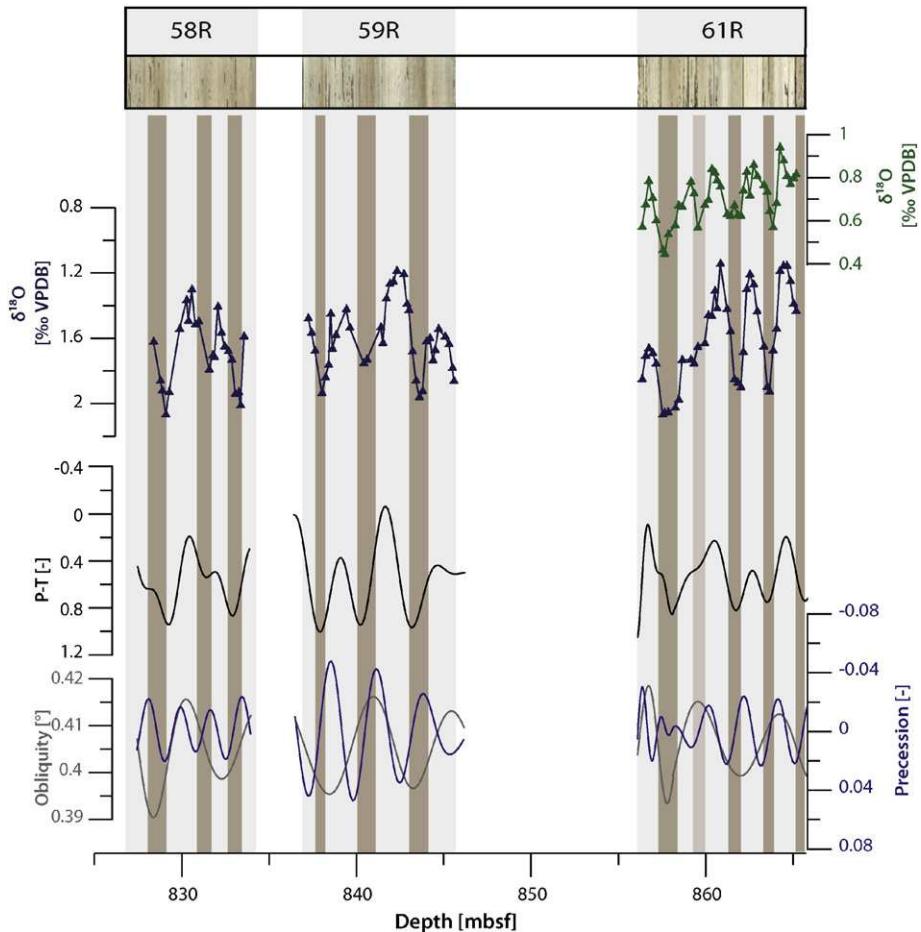


Fig. 9. Close-up of $\delta^{18}\text{O}$ (dark blue) and $\delta^{13}\text{C}$ (green) of *C. pachyderma* together with colour changes from 827.20 to 865.85 mbsf (Cores 58R, 59R and 61R). A theoretical example of a corresponding precession-tilt curve (P-T; black) is displayed to show resemblance with $\delta^{18}\text{O}$. This suggests the interference patterns of precession and obliquity in the $\delta^{18}\text{O}$. The used obliquity (grey) and precession (light blue) curves are indicated below. Parts of the solution La04 (65°N summer insolation; [Laskar et al., 2004](#)) are used for all astronomical parameters.

(Interval I) of IODP Hole U1387C. This, combined with the relatively low Zr/Al ratios, which shows no clear correlation with the precessional benthic $\delta^{18}\text{O}$ record, indicate low energy depositional environment influenced by orbital variation (Bahr et al., 2014). Similar cyclicity is found in NW Morocco and the southern Iberian Margin, where clay-silt alternations result from orbitally-driven oscillations in detrital input (coast, riverine, or eolian; e.g. Sierro et al., 2000; van der Laan et al., 2012; Van den Berg et al., 2015). Above the Miocene–Pliocene boundary (Interval II and III), sedimentation rates are about three times higher than during the Messinian and the succession has higher Zr/Al values and sand content with more abundant lithic particles (Fig. 3). This suggests significant change in the depositional environment from quiet hemipelagic conditions to a higher energy environment including the presence of weak bottom currents.

6.1. Bottom water flow strength indicators

Zr/Al ratios are relatively high with large amplitude variations in Interval II and III. The sudden change in the appearance of the Zr/Al record is therefore interpreted as a shift from a low energy hemipelagic setting in Interval I, to a higher energy environment controlled by bottom currents that are subject to particle sorting in Interval II and III (Fig. 3). The suggestion of increasing energy levels is consistent with the general increase in sand content in Intervals II and III relative to the base of the Hole, especially in sections where maximum Zr/Al ratios coincide with maximum sand content (i.e. 788.80 to 805.68 mbsf; Fig. 4). Yet because of the absence of Zr/Al data from the lowermost core of Interval II (Core

57R), we cannot determine with certainty from Zr/Al ratios only whether this transition occurs at the Miocene–Pliocene boundary or shortly thereafter. However, parallel behaviour between Zr/Al ratios and $\delta^{18}\text{O}$ immediately after the oldest measurement in Interval II and the abrupt sedimentation rate change reflected in benthic $\delta^{18}\text{O}$, suggests an immediate change of depositional environment at the Miocene–Pliocene boundary.

In Interval II, two sandy beds show clear bigradational grading with tracers of bioturbation (Core 54R and 55R; 788.88 to 803.68 mbsf; Figs. 2C, D, 4). The patterns represent a coarsening upward sorting from homogenous muds with sparse bioturbation to more silty to sandy deposits until the facies succession reverses. These features are characteristic of contourites (Faugères et al., 1984; Gonthier et al., 1984; Rebasco et al., 2014). Similar sedimentary sequences to the two sandy beds are characterized as contourite sequences in the Faro Drift (Gonthier et al., 1984) and recently envisioned for the early Pleistocene for Site IODP U1386 in Alonso et al. (2016). These are distinctly different from the instantly deposited (un-bioturbated) turbiditic, normal grading, fining upward sequences (cf. Bouma et al., 1962). Discrimination between the sand grades of >63 to $<150\text{ }\mu\text{m}$ and $>150\text{ }\mu\text{m}$ weight % confirms the hypothesis of continuous sedimentary deposition under the influence of bottom water current flow. Therefore, the two sandy beds with bi-directional grading are interpreted as the earliest clear sandy contouritic deposits above the third precession cycle above the tentative Miocene–Pliocene boundary.

Seismic data imaging of the early Pliocene shows features of slope instability on the Algarve Margin, north of IODP Site U1387

(Brackenridge et al., 2013; Hernández-Molina et al., 2014b, 2016; Martínez del Olmo, 2004; Roque et al., 2012). This sedimentary source may account for the tripling of the sedimentation rate that occurs at the Miocene–Pliocene boundary. However, the change in sedimentation rate is abrupt, rather than gradational, which is what would be expected if sedimentation rate changes as a result of progressive tectonic uplift of the margins and/or infilling of the basin, such as is seen in the Guadaluquivir Basin (Van den Berg et al., 2015). There is also no evidence of significant shallowing of the depositional environment in Hole U1387C during this period. The continuous presence of benthic foraminiferal taxa which have a bathyal upper depth limit, suggest a fairly constant bathymetry throughout Intervals I to III. The increasing presence of benthic foraminiferal species limited to neritic environments from about 801 mbsf (i.e., in Interval II, ~25 m above the Miocene–Pliocene boundary) upwards are interpreted as allochthonous. They always occur alongside species with a bathyal upper depth limit and are thus not indicative of a significant shallowing of the water depth at Hole U1387C. The shallower benthic foraminifers are probably introduced by pulses of episodic turbiditic-like downslope transport, resulting in the cm-scale sandy layers on the upper slope (Fig. 2D; 'sand influx' Fig. 4). In most cases, these thin turbiditic beds are almost destroyed by the combined action of bottom currents and macrobenthic communities that buried and excavate within the sediments (Dorador and Rodríguez-Tovar, 2016; Rodríguez-Tovar et al., 2015). Often only small patches of sand were preserved within a matrix of mud (Fig. 2F). If the increase in sedimentation rate was the result of instability on the Algarve Margin, then this shallow-water material likely resulted from direct downslope sediment transport. If, however, the abrupt sedimentation rate increase resulted from the onset or an increase of along-slope transport, then the shallow water material could derive from an up-current source. Most likely, a combination of the two processes has occurred; increased slope instability due to tectonic uplift causing increased sediment input from the northern margin, while along-slope flow provided suspended load from the distant Guadaluquivir drainage basin, similar to the late Pleistocene described for IODP Sites U1387 and U1386 (Alonso et al., 2016). In addition, the along-slope suspended load may have been amplified due to suggested tectonic uplift from the early Pliocene (Hernández-Molina et al., 2014b). Interval III shows even higher grain size fractions (>63 µm) than Interval II, suggesting increased influence of the along- and/or downslope processes. The possible increased terrigenous input from the Guadaluquivir and Guadiana rivers into the Gulf of Cadiz during the warm climate of the early Pliocene (Ducassou et al., 2016; Miller et al., 2005) may have amplified the along and/or downslope processes.

In summary, Hole U1387C shows features in Intervals II and III (lowest Pliocene) that are consistent with transport and deposition by weak along-slope bottom water currents. These features are not present in Interval I (upper Miocene), that suggests low or absent influence of bottom current flow across Site U1387. Thus, either an increase or the onset of the bottom water current occurred immediately at or shortly after the Miocene–Pliocene boundary. The overall increase in sand content in Intervals II and III may reflect a further increase in flow strength during the earliest Pliocene. Likely, the increase in sedimentation rate at the Miocene–Pliocene boundary is the result of a combination of local instability of the margins and the onset of along-slope transport.

6.2. Origin of bottom water current flow

MOW today is apparent as a warmer intermediate water mass along the margins of south west Iberia penetrating the Gulf of Cadiz from the Gibraltar Strait (Hernández-Molina et al., 2014a). Consequently, if the Strait were closed, colder intermediate Atlantic waters would be expected at the same water depths in the Gulf of Cadiz. The temperature effect should be reflected in the benthic foraminiferal $\delta^{18}\text{O}$, where colder conditions and decreased MOW result in an increase in the $\delta^{18}\text{O}$ values. The benthic stable isotope records of NE Atlantic ODP Site

982 (1134 m water depth; Hodell et al., 2001) and onshore Ain el Beida (Krijgsman et al., 2004; van der Laan et al., 2005) and Loulja sections (Van der Laan et al., 2006) are compared with the benthic stable isotope record of Site U1387 in order to evaluate shorter scale patterns and relative off-sets (Fig. 8). The offset of about 1.2 ‰ between the benthic $\delta^{18}\text{O}$ records from ODP Site 982 and NW Morocco has been interpreted as the temperature difference between cold Upper North East Atlantic Deep Water (UNEADW) at Site ODP 982 and warmer near-surface waters in NW Morocco in Van den Berg et al. (2015). The offset between the U1387C benthic oxygen isotope record and ODP 982 is smaller than its offset relative to NW Morocco, particularly during the Messinian (Fig. 8B). Ignoring the potential changes in the $\delta^{18}\text{O}$ of the water, the reduced $\delta^{18}\text{O}$ -offset between Site ODP 982 and IODP U1387 can indicate that temperature of intermediate Atlantic waters in the Gulf of Cadiz was closer to that of UNEADW in the late Miocene. Counterintuitively, the $\delta^{18}\text{O}$ of today's MOW is higher (0.5 to 1.1 ‰) compared to NEADW (−0.1 to 0.5 ‰), however, the temperature difference between these two water masses is between 8 and 10 °C (Voelker et al., 2015a). Hence the temperature effect on the shell of the benthic foraminifer $\delta^{18}\text{O}$ is much higher than the effect of $\delta^{18}\text{O}$ of the water (Epstein et al., 1951, 1953). This relation is evident during the transition from Marine Isotope Stage 32 to 31 in IODP Hole U1387, where an increase of MOW (high XRF scanning Zr/Al ratios) results in lower benthic foraminiferal $\delta^{18}\text{O}$ (Voelker et al., 2015b). Therefore, the observed decrease in benthic $\delta^{18}\text{O}$ during the early Pliocene and its stepped divergence away from ODP 982 values suggests a gradual warming of intermediate Atlantic Waters near Gibraltar. This temperature increase over the studied period can be attributed to the absence or minor influence of warm MOW during the Messinian and the presence or increase of MOW during the Pliocene. Furthermore, increasing amplitudes that appear in Interval II are even larger in Interval III and may be attributed to increasingly fluctuating temperatures caused by varying influences of MOW over periods of insolation minima and maxima. Both arguments suggest that the increase or onset of bottom water current flow over the Miocene–Pliocene boundary can be attributed to MOW. A significant shallowing that would cause the change in bottom water temperatures at Site U1387C during the earliest Pliocene is excluded based on the benthic foraminiferal analyses.

Interestingly, amplitudes of benthic $\delta^{18}\text{O}$ at Site U1387 do not only increase over the Miocene–Pliocene boundary, but also during the Messinian they are larger than amplitudes at Site ODP 982, Ain el Beida and Loulja. The Montemayor-1 core reveals the same feature as Hole U1387C (Pérez-Asensio et al., 2012; Van den Berg et al., 2015). The difference in amplitude of Hole U1387C and ODP 982 must be attributed to either a relative change of $\delta^{18}\text{O}$ in the water mass or temperature. Increased amplitudes during the Pliocene can be attributed to fluctuations in MOW, however, it is thought that during the MS C MOW was either extensively reduced or even absent (Flecker et al., 2015). Larger amplitudes due to larger temperature changes of shallower water depths can be excluded, because the shallower Loulja and Ain el Beida sections do not show the larger amplitudes. These sections are located closer to the thermocline and thus more sensitive to temperature changes on glacial–interglacial cycles. One alternative, however speculative since there are no flow strength indicators observed in the sediments, is that the relatively high amplitude variations in benthic $\delta^{18}\text{O}$ in U1387C compared with both shallower and deeper locations hint towards the presence of weak MOW during the Messinian.

7. Conclusions

We present evidence that the flow of Mediterranean–Atlantic bottom water began very shortly after the Messinian Salinity Crisis in the Gulf of Cadiz from IODP Site U1387. This study provides a chronostratigraphic framework for the lower part of IODP Site U1387C utilizing biostratigraphic, paleomagnetic, and cyclostratigraphic constraints. Our data suggest that the record extends into the upper Messinian and

includes the Miocene–Pliocene boundary. Seismic data linking Site U1387C and the Algarve-2 well, as well as sedimentary changes in the sequence itself constrains the Miocene–Pliocene boundary at about 826 mbsf. Sediment deposition during the latest Messinian in the Gulf of Cadiz took place in a relatively quiet, hemipelagic environment driven by precessionally controlled Mediterranean monsoonal induced oscillations, with negligible evidence of bottom water flow strength. During the earliest Pliocene, an immediate decrease in the benthic $\delta^{18}\text{O}$ may be the result from the direct warming of intermediate Atlantic waters due to increased Mediterranean–Atlantic exchange. The increase in sedimentation rate and as well as elevated variable XRF scanning Zr/Al ratios could have resulted from the onset of weak along-slope bottom water currents immediately at or right after the Miocene–Pliocene boundary. The occurrence of contouritic sedimentation in the Gulf of Cadiz results in two clear bigradational sandy beds above the third precession cycles after the Miocene–Pliocene boundary. Shallow water benthic foraminifers indicate that, besides along-slope transport, downslope transport was present, redistributing shallow marine (continental shelf) particles to the upper slope.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <http://dx.doi.org/10.1016/j.margeo.2016.04.005>.

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