No major deglaciation across the Miocene-Pliocene boundary: Integrated stratigraphy and astronomical tuning of the Loulja sections (Bou Regreg area, NW Morocco)

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[1] An integrated high-resolution stratigraphy and orbital tuning is presented for the Loulja sections located in the Bou Regreg area on the Atlantic side of Morocco. The sections constitute the upward continuation of the upper Messinian Ain el Beida section and contain a well-exposed, continuous record of the interval straddling the Miocene-Pliocene (M-P) boundary. The older Loulja-A section, which covers the interval from ~5.59 to 5.12 Ma, reveals a dominantly precession-controlled color cyclicity that allows for a straightforward orbital tuning of the boundary interval and for detailed cyclostratigraphic correlations to the Mediterranean; the high-resolution and high- quality benthic isotope record allows us to trace the dominantly obliquity-controlled glacial history. Our results reveal that the M-P boundary coincides with a minor, partly precession-related shift to lighter “interglacial” values in d18O. This shift and hence the M-P boundary may not correlate with isotope stage TG5, as previously thought, but with an extra (weak) obliquity-controlled cycle between TG7 and TG5. Consequently, the M-P boundary and basal Pliocene flooding of the Mediterranean following the Messinian salinity crisis are not associated with a major deglaciation and glacio-eustatic sea level rise, indicating that other factors, such as tectonics, must have played a fundamental role. On the other hand, the onset of the Upper Evaporites in the Mediterranean marked by hyposaline conditions coincides with the major deglaciation step between marine isotope stage TG12 and TG11, suggesting that the associated sea level rise is at least partly responsible for the apparent onset of intermittently restricted marine conditions following the main desiccation phase. Finally, the Loulja-A section would represent an excellent auxiliary boundary stratotype for the M-P boundary as formally defined at the base of the Trubi marls in the Eraclea Minoa section on Sicily.


1. Introduction

[2] The Miocene-Pliocene boundary formally defined at the base of the deep marine Trubi marls on Sicily [Van Couvering et al., 2000] marks the basal Pliocene flooding of the Mediterranean and the end of the Messinian salinity crisis (MSC). Although there are increasing indications that (restricted) marine conditions may have occurred already intermittently during deposition of the Upper Evaporites [e.g., Müller, 1990; Castradori, 1998; Spezzaferri et al., 1998; Iaccarino et al., 1999; Rouchy et al., 2001; Aguirre and Sanchez-Almazo, 2004], the base of the Trubi marls and time equivalent sediments elsewhere in the Mediterranean represent(s) the sedimentary expression of the reestablishment of full open marine conditions in the Mediterranean [Hsiü et al., 1973; Cita, 1975; Hilgen and Langereis, 1993; Iaccarino et al., 1999]. To unravel the exact cause of the flooding event, the time-correlative level of the M-P boundary should be precisely pinpointed in a continuous orbitally tuned open marine succession outside the Mediterranean. In addition an excellent benthic oxygen isotope record should be available to detect any significant reduction in ice volume associated with the flooding event.

[3] Attempts have recently been made to pinpoint the position of the M-P boundary in marine cores from the open ocean. Shackleton et al. [1995a, 1995b], Clauzon et al. [1996], and McKenzie et al. [1999] all link the boundary to the prominent interglacial oxygen isotope stage TG5 defined by Shackleton et al. [1995b] in ODP Site 846, suggesting that the associated glacio-eustatic sea level rise played a primary role in controlling the Pliocene flooding of the Mediterranean. For this purpose most of these studies compared the tuned age of 5.33 Ma for the base of the Trubi [Lourens et al., 1996] with the age of TG5 according to the initial tuned age model developed for Site 846 [Shackleton et al., 1995a] but the latter age model has subsequently been revised leaving a younger age of 5.31 Ma for TG5 [Shackleton et al., 1999].

[4] In our opinion, the Bou Regreg area located on the Atlantic side of Morocco remains the location most critical to solve the problem of the exact timing and origin of the
Pliocene flooding of the Mediterranean. The open marine succession in the area shows a distinct dominantly precession-controlled color banding that is related to the sedimentary cyclicity in the Mediterranean, and the Trubi marls in particular. The color cycles allow the sections to be astronomically tuned and to be correlated cyclostratigraphically to Mediterranean sections [Hilgen et al., 2000]. Secondly, high-quality benthic isotope records can be generated due to the good to excellent preservation of the foraminiferal shells while the relatively high average sedimentation rate (of 7–8 cm/kyr) guarantees an optimal temporal resolution [Van der Laan et al., 2005].

In fact, Benson and Rakic-El Bied [1996] proposed to define the M-P boundary in a continuous open marine section, Ain el Beida, located outside the Mediterranean because they considered the Messinian to be a regional stage representing the sedimentary expression of the MSC only. For this purpose, they selected a much older level that corresponds with the younger end of Chron C3An. On the other hand, Sue et al. [1997], while recognizing the problem of the possible unconformable nature of the base of the Trubi, preferred to maintain the Mediterranean option but proposed to define the M-P boundary near Salé, some 5 km northwest of Ain el Beida, at the stratigraphic level that corresponds with the base of oxygen isotope stage TG5. This isotope shift was supposed to reflect the marked deglaciation caused responsible for the Pliocene flooding of the Mediterranean. The proposal was considered less favorable because the level was only indicated in the Salé drill core but not in any existing outcrop.

Nevertheless, the M-P boundary level might be present in one of the quarries located in between Ain el Beida, where this level is not reached, and Salé with its exposures of early Pliocene marine sediments. On the basis of the clearly visible color banding we selected the Loulja section with the aim (1) to extend the tuned open marine succession exposed at Ain el Beida across the M-P boundary into the Pliocene and (2) to pinpoint as accurately as possible the position of the M-P boundary in a well-tuned open marine section with an excellent oxygen isotope stratigraphic record in the Bou Regreg area. The outcome of this study may provide the final answer to the question whether glacio-eustatic sea level rise was involved in triggering the Pliocene flooding of the Mediterranean.

In addition, special attention will be given to identify the major isotope shift to lighter values between isotope stages TG12 and 9. According to the tuned age model for Ain el Beida, only the beginning of this shift was reached in the topmost part of the section [Van der Laan et al., 2005]. This shift, which marks a more dramatic deglaciation than TG5 as noticed by Shackleton et al. [1995b], Hodell et al. [2001], and Vidal et al. [2002], is particularly interesting because it may be coincident with and thus responsible for the beginning of the Upper Evaporites in the Mediterranean [Shackleton et al., 1995b].

## 2. Setting and Sections

### 2.1. Bou Regreg Valley

The classic Blue Marl of Atlantic Morocco has been subject of numerous studies, reflecting the progress made in Neogene stratigraphy. It is best known from the Bou Regreg river valley where it constitutes a continuous open marine record from the late Tortonian into the Pliocene that overlies Devonian limestones with an angular unconformity [Benson and Rakic-El Bied, 1996]. The Blue Marl was deposited in the Gharb basin, which formed the westward extension of the Rfian Corridor (Figure 1a). This corridor acted as an extensional foredeep and formed one of the two Atlantic-Mediterranean connections during the late Miocene before its tectonic closure in the course of the Messinian [Benson et al., 1991; Krijgsman et al., 1999]. Deposition of the Blue Marl in the Gharb Basin continued well into the Pliocene and its stratigraphic succession is exposed in a number of quarries and outcrops albeit in an incomplete way. The Oued Akrech section contains the basal part of the Blue Marl, straddling the Pliocene/Messinian boundary and covering the interval from 7.6 to 7.1 Ma [Hilgen et al., 2000]. The middle part of the Blue Marl is exposed in the Ain el Beida quarry and covers the interval from 6.5 to 5.5 Ma [Krijgsman et al., 2004], while lower Pliocene marine sediments are exposed at Salé [see Benson and Rakic-El Bied, 1996; Hodell et al., 1994]. The succession has been completed by drilling at Ain el Beida and Salé [Hodell et al., 1989; Benson et al., 1991; Hodell et al., 1994; Benson et al., 1995], but up to now, the interval straddling the Miocene-Pliocene boundary has not been documented in detail from any outcrop.

### 2.2. Loulja-A Section

The Loulja-A section is located in a brick quarry at the northern side of the Bou Regreg valley along the road to Salé (Figures 1a and 1b). Color banding is clearly visible in the steep south face of an old quarry next to the road, which shows a regular alternation of slightly protruding, light beige colored silty marls and softer, darker (reddish) colored, clayey marls (partial section I, Figure 1c). A much larger quarry is located on the opposite northern side of the hill. The weathered south face (partial section II, Figure 1d) of this quarry shows the same color banding and pattern as observed along the road. The entire sequence is also exposed in the relatively fresh, north face (partial section

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**Figure 1.** (a) Simplified geological map of the Gibraltar area with the location of the late Miocene Betic and Rifian Corridors (modified after Krijgsman et al. [2004]). (b) Map of the lower Bou Regreg valley showing locations of the quarries of Loulja-A and Loulja-B, Ain el Beida and Salé Briqueterie, and the road section of Oued Akrech (modified after Millies-Lacroix [1974]). (c) Old quarry of Loulja-A (partial section I) along the road to Salé. (d) View in NW direction, toward the Loulja-B and Salé quarries showing the weathered side (section II) of the main Loulja-A quarry. White horizontal bar in inset shows detail of sample trajectory. (e) Fresh north facing side (section III) side of the main Loulja-A quarry. White bars indicate sample trajectories. (f) Detail of right, southeast facing wall of the Loulja-B quarry (inset shows the entire quarry). Note the variable meter scales in Figures 1c–1f. Cycle numbers refer to the stratigraphy in Figure 2.
Figure 1
III, Figure 1e) of the quarry, which is approximately 60 m high. Younger stratigraphic levels showing relatively thick color cycles are reached in the upper half of this quarry face. The color cycles are less distinct in such fresh exposures and consist of alternating lighter and darker colored bluish-gray marls.

[10] Sampling started at the base of the accessible eastern side of section Loujia-A III (lower 20 m) and was continued along the path that crosses section II (from 9 to 33.5 m). An additional 5.5 m was drilled below the reference level of the quarry floor with a hand drill at section III (Figure 1e) to ensure a stratigraphic overlap with the top part of the (supposedly older) Ain el Beida section. Usually 8 to 10 samples were taken per cycle, which corresponds to a temporal resolution of 2–3 kyr in case the color cycles are precession related. In the drill hole, one sample was taken every 10 cm. Finally, additional samples with a resolution of two samples per cycle were taken from the lower part of section I and from the lower part of section II for a check on the cycle patterns and the correlations between partial sections. Sampling was not extended to include all the thick cycles observed in the top part of Loujia-A due to the intense weathering of section II and the steepness of the upper part of section III.

2.3. Loujia-B Section

[11] The Loujia-A section can be extended upward in another brick quarry along the road to Salé (Figures 1b and 1f), located about 1.5 km northwest of Loujia-A. This section, named Loujia-B, displays similar thick and well-developed color cycles as the upper half of the Loujia-A section. Inaccessibility of the vertical southwest face of this quarry limited the detailed sampling to the lower half of the exposed section. Nevertheless, seven additional samples were collected from the upper four cycles for biostratigraphic analysis.

3. Cyclostratigraphy

[12] In total 27 cycles were recognized at Loujia-A, labeled LA-1 to LA-27 in stratigraphic order (Figure 2a). In weathered outcrop the cycles consist of an alternation of more indurated light beige colored silty marls and softer reddish colored clayey marls. Rather arbitrarily, reddish marl layers are taken as to define the base of a cycle and labeled LA-X_B where X is the number of the cycle in the section and B denotes the color of the marl (red); the overlying beige marl is denoted LA-X_S (as in Figure 3b), after the overlying beige marl is denoted LA-X_B. The lowermost 3 cycles are found only in the drill hole and are distinguished on the basis of the geochemical data (see section 6). Cycles LA-4 to base LA-21 were logged and sampled in detail in partial sections II and III whereas only the (approximate) thickness of cycles LA-21 to LA-27 was measured with a tape from the top of section III downward.

[13] The cycles reveal a very characteristic pattern (Figure 2a). LA-1 to LA-10 reach an average thickness of 2 m with very thick reddish layers in LA-7 and LA-10 and relatively thin reddish layers in LA-5 and LA-6. The thickest whitish layers are found in LA-1 and LA-9. The lower part of the section is followed by a very characteristic interval, which consists of 6 successive thin color cycles (LA-10 to LA-15), which are very regular and have an approximate thickness of 1 m. This interval can be recognized both in partial section I as well as in sections II and III, and was used to correlate the different partial sections in detail. Cycle LA-16 has an intermediate thickness and marks the transition to thicker cycles LA-17 to LA-21 (and up to LA-27). The thickness of these cycles ranges from 2.5 to 4 m.

[14] The Loujia-B section contains 10 thick color cycles labeled LB-1 to 10 (Figure 2b). The lowermost 7 cycles have an average thickness of ~4 m with a very thick reddish layer of 5 m in LB-2. Cycles LB-8 and LB-9 are extremely thick and reach a thickness of 7.5–8 m, i.e., almost twice as thick as LB-1 to 7.

4. Magnetostratigraphy

4.1. Sampling

[15] Standard oriented paleomagnetic cores were drilled with the help of a water-cooled, generator-powered electric drill at two to twelve levels per sedimentary cycle (four on average). Samples were taken in partial sections I and II (between –1 and 33 m; Figures 1c, 1d, and 2a). Care was taken to remove the weathered surface before drilling. In addition to the regular sampling, oriented hand samples were collected from the middle part (10–20 m) of section III with a resolution of one sample per cycle. Standard cores were drilled from these samples with compressed air at the paleomagnetic laboratory Fort Hoofddijk.

4.2. Demagnetization

[16] One specimen per sampling level was thermally demagnetized with temperature increments of 15°–80°C up to a maximum temperature of 700°C in a magnetically shielded, laboratory-built furnace. The natural remanent magnetization (NRM) of the specimens was measured on a 2G Enterprises horizontal DC SQUID cryogenic magnetometer.

[17] Demagnetization diagrams [Zijderveld, 1967] are of mixed quality (Figure 3). A normal polarity component is generally removed at temperatures between 100 and 250°C (e.g., Figure 3b). This relatively low-temperature component typically has a present-day field direction and is interpreted as a secondary, recent overprint, likely caused by weathering. Therefore low-intensity samples showing clustering from 250°C onward are ignored. A relatively high temperature component is gradually removed between temperatures of 280 and 360°C (as in Figure 3d), after which spurious magnetizations prevail, probably due to the formation of pyrite or magnetite. In the last group of samples, the high-temperature component is removed between 280 and 600°C, up to 700°C (e.g., Figure 3b), indicating magnetite or haematite as the principal carriers of the magnetization. The high-temperature component is taken as the characteristic remanent magnetization (ChRM) direction. In approximately 50% of the samples, the demagnetization diagrams allow straightforward interpretation of reversed (Figures 3a, 3b, and 3e) or normal (Figures 3c and 3d) polarity. In other samples, (linear) decay toward the reversed side of the diagram occurs (Figures 3f–3h). In these cases, great circles are used to approximate the reversed ChRM
Figure 2. Lithostratigraphy, biostratigraphy, and magnetostratigraphy of the (a) Loulja-A and (b) Loulja-B sections. Sedimentary cycle numbers are based on the color cycles and, for the lowermost part of Loulja-A, also on the PC-1 data shown in Figure 4. Light shading marks the light colored, relatively carbonate-rich marls, and darker shading marks the relatively clayey, often reddish colored marls. Last and first (regular) occurrences of nannofossil species are indicated on the left side of the lithological columns: D.q., Discoaster quinqueramus; C.l., Ceratolithus larrymayeri; Tr., Triquetrorhabdulus rugosus; A.t., Amaurolithus tricorniculatus; A.b., Amaurolithus bizzarus; R.p., Reticulofenestra pseudoumbilicus (>8 µm); D.p., Dictyococcites perplexus; P.l.o., Pseudoemiliania lacunosa ovata; H.s., Helicosphaera sellii. Furthermore, G.m.-interval stands for the interval with common Globorotalia menardii. In the VGP curves, solid (open) circles denote (less) reliable ChRM directions; asterisks indicate great circle solutions; pluses are used for inconclusive results. In the polarity columns, black (white) denotes normal (reversed) polarity, gray indicates undetermined polarity, and crossed intervals are used for those parts of the sections where no paleomagnetic samples were taken. Magnetic susceptibility is shown to the right of the VGP curves; II and III refer to partial sections of Loulja-A (see Figure 1). Stratigraphic scales are in meters.
Figure 3
directions. Best fitting directions from the great circles (indicated by small circles in Figures 3i–3l) are determined by using the reliable directions as set points (thick circles) according to the method of McFadden and McElhinny [1988].

[18] The ChRM directions, represented as virtual geomagnetic polar (VGP) latitudes, reveal two intervals of normal polarity: from 20.53 ± 0.18 m upward in the Loulja-A section (LA-16 to LA-21; Figure 2a), and below 10.70 ± 0.90 m in cycle LB-3 in the Loulja-B section (Figure 2b).

4.3. Magnetic Susceptibility

[19] The initial bulk magnetic susceptibility (χm) was measured at room temperature on a KLY-2 Kappabridge [Hrouda, 1994]. Magnetic susceptibility (MS) is a measure of the magnetic response of a sample to a small, applied magnetic field and proportional to the concentration and grain size of ferromagnetic and paramagnetic minerals in the sample. It can therefore be used to detect variations in lithology.

[20] Magnetic susceptibility values that are not consistent with the lithological expression of the corresponding color cycle (LA-7).

[21] By contrast, MS values are much lower in the younger cycles (LA-12 to LA-15 up to 18.15 m in section III) while changes are less pronounced. Also the records obtained from the basal part of section III (−5.5 to −2 m), from section II (14 to 33.5 m) and from Loulja-B (0−23 m) show low values of MS, not exceeding 3−6 × 10^{-6} SI units (Figures 2a and 2b).

5. Biostratigraphy

5.1. Planktonic Foraminifera

[22] A semi-quantitative analysis of potential marker species was carried out on the >125 μm fraction of the washed residue of at least every other sample. In addition, the coiling direction of the neogloboquadrinids was determined. Sphaeroidinellopsis spp. and Globorotalia margaritae are present throughout the section and do not reveal changes in abundance that allow recognition of the Sphaeroidinellopsis acme and the Globorotalia margaritae FCO known from the Mediterranean. Next to G. margaritae, another species of left coiling keeled globorotalids, G. menardii, is commonly but discontinuously found over a short interval in the lower part of the section (from −1.75 ± 0.20 m up to 2.15 ± 0.05 m, Figure 2a). Its distribution corresponds to the reappearance of common G. miotumida group in the top part of the underlying Ain el Beida section [Krijgsman et al., 2004].

[23] The determination of the neogloboquadrinid coiling direction is complicated by the presence of two different types, Neogloboquadrina acostaensis and N. humerosa. The coiling direction of the small- to medium-sized N. acostaensis type is dominantly dextral. Approximately equal numbers of sinistral and dextral specimens are found at the 14.65 m level (LA-11b) and a short but prominent incursion of left-coiling specimens occurs between 16.15 and 16.75 m (cycle LA-13; see also Figure 2a). Specimens of the N. humerosa type, which are often large-sized, are frequently encountered in the lower part of the section although they are always less abundant than N. acostaensis. Their coiling direction is dominantly sinistral. Dominantly dextral N. acostaensis is the only neogloboquadrinid species present from the top of the marked (second) sinistral influx upward.

[24] An estimate of the paleobathymetry of the Loulja sections using the ratio of planktonic and benthic foraminifers [Van Hinsbergen et al., 2005] yielded values between 300 and 500 m. A check on this ratio for the Ain el Beida section revealed paleobathymetric values between about 600 and 1000 m, which are in good agreement with previous estimates [Benson and Rakic-El Bied, 1996].

5.2. Calcareous Nannofossils

[25] The samples selected for analysis of the calcareous nannofossil content were prepared according to standard preparation techniques [Bramlette and Sullivan, 1961]. Two samples per sedimentary cycle were qualitatively examined from both Loulja-A and Loulja-B. Samples from Loulja-A III (−5.5−21 m) provided the best preserved nannofossil assemblages; all other samples yielded moderately to well-preserved nannofossils.

[26] The Loulja-A section comprises several important nannofossil bioevents (Figure 2). Only a single specimen of Discoaster quinqueramus was observed at −0.7 m in cycle LA-3, pinpointing the LO of the species between −0.7 m and −0.18 m (Figure 2a). Rare specimens of Ceratolithus larrymayeri were found between 15.5 m and 17.9 m (cycles LA-12 to LA-14), defining the FO and LO of this species. The Triquetrorhabdus rugosus LO was reached between 18.2 and 18.3 m (LA-15), while Amaurolithus

Figure 3. Examples of stepwise thermal demagnetization of selected samples. Diagrams denote orthogonal projections of NRM vector end points, where open (solid) symbols are used for the vertical (horizontal) plane. Values represent temperatures in °C; stratigraphic level is indicated in the lower left corner. Reliable samples show both (a, b, e) reversed and (c, d) normal polarity. (f–h) For samples where the ChRM is less well resolved, great circles used to approximate the (reversed) directions, with squares indicating the normals to the great circle planes and ellipses representing the maximum angular deviations. (i–l) Best fitting directions from the great circles shown per analyzed interval in the Loulja-A and Loulja-B sections (small circles). Thick circles indicate the used set points; small and large ellipses denote the χ_{0.95} and χ_{0.95} confidence limits [McFadden and McElhinny, 1988].
tricorniculatus was first observed at 30.3 m (LA-20; A. tricorniculatus FO between 28.0 and 30.3 m).

[27] In Loulja-B, Amaurolithus bizzarus was found at the 5 m level (LB-2, A. bizzarus FO between 2.25 and 5.0 m in Figure 2b). The subbottom of Reticulofenestra pseudoumbilicus (>8 μm) is recorded between 17.8 m and 19.4 m, in cycle LB-5 while the Dictyococcosites perplexus subtop is placed between 19.4 m and 21.7 m (LB-5g and LB-6g) and the Pseudoemiliania lacunosa ovata FO (forms with less than 12 rim slits) between 21.7 m and 22.5 m, in cycle LB-6; Finally, Helicosphaera sellii first appeared between 21.7 and 22.5 m (LB-6; H. sellii FO), and is more frequently present from between 25 and 27 m upward (LB-7; H. sellii FRO).

6. Chemostratigraphy

6.1. Geochemistry

[28] Samples were dried, crushed, powdered and homogenized in an agate mortar prior to further treatment. Approximately 125 mg of each sample was then dissolved in 2.5 ml HF (40%) and 2.5 ml of mixed HNO₃ (16.25%) and HClO₄ (45.5%) and heated at 90°C in closed “Teflon” tubes (bombs) for a minimum of 8 hours. The solutions were subsequently evaporated (in open bombs) at a temperature of 160°C, after which approximately 25 ml HNO₃ (4.5%) was added to each residue. The resulting solutions were analyzed using a Perkin Elmer Optima 3000 ICP-OES apparatus for the elements Al, Ba, Ce, Ca, Co, Cr, Cu, Fe, K, Li, Mg, Mn, Na, Ni, P, Sc, Sr, Ti, V, Y and Zn. The relative errors in duplicate measurements of international standards was lower than 3% for all elements, except for Ce, Co, P and S, Sr and Y (>5%).

[29] A standardized Principal Component Analysis (PCA) (SPSS software) was applied to extract the primary components responsible for the main variance in the data set. The first (PC1) and second principal component together describe already 73% of the total variance in the data, of which PC1 explains about 59% and is statistically significant. The interpretation of the first principal component is based on the loadings of individual elements on this component (Table 1). PC1 reveals very high positive loadings for K, Al, V, Fe, Ti, and Li whereas Ca displays the strongest negative loading. The elements that show high positive loadings are main constituents in clay minerals and often represented in the fine-grained (clayey) terrigenous sediment fraction. On the contrary, Ca is primarily associated with the marine (biological) fraction of the sediment. The PC1 record shows obvious cyclic variations and follows the sedimentary color cycles observed in the field, where PC1 minima and Ca maxima correspond to the more indurated, beige colored marls. The record comprises 21 cycles that vary in thickness between about 1 and 4 m with thicker cycles generally showing higher-amplitude changes (Figure 4). Thicker geochemical cycles are found between −4.5 and +2.5 m (which correspond to color cycles LA-2 to LA-4) and between 5.5 and 8 m (LA-7) in the lower part of the section and between ~21 m and 33.5 m (cycles 17–21) in the upper part. The middle part reveals a series of six thin geochemical cycles with low-amplitude changes (Figure 4). This part, which is marked by a decrease in PC1 values, corresponds to the characteristic interval with thin color cycles observed in the field (cycles LA-10b-15b: Figures 1c and 2a).

6.2. Stable Isotopes

[30] For isotope analyses, between 15 and 40 specimens (size-dependent) of the benthic foraminifera Planulina ariminensis were picked from in total 145 samples. Preservation is generally good to excellent without any indication for recrystallization or pyritization. Where P. ariminensis was not abundant enough (<500 μg per sample), the Cibicides species C. dutemplei and C. pachyderma were picked in addition. In order to remove any organic remains, each sample was roasted for 30 minutes at 470°C under vacuum. The samples were analyzed using an ISOCARB, which is directly coupled to the mass spectrometer and has the capacity to measure 44 samples, including 1 international (IAEA-CO-1) and 9 in-house (NAXOS) standards, during a run. Each sample reacted with 103% phosphoric acid (H₃PO₄) for 6 to 7 minutes at 90°C. All isotope data are reported as per mil (%o) relative to the PeeDee belemnite (PDB) standard. After correction of the Cibicides values by adopting a value equal to the average offset from the Planulina values, twelve data points were incorporated in the δ¹⁸O series. The analytical precision and accuracy were determined by replicate analyses of samples and by comparison with the IAEA-CO-1 standard. The relative standard deviations, analytical precision and accuracy were better than 0.1‰ both for δ¹⁸O as well as for δ¹³C.

6.2.1. The δ¹⁸O Record

[31] The δ¹⁸O record shows a stepwise shift in mean δ¹⁸O to lighter values from −5.5 to +8 m. The average δ¹⁸O value is 1.05‰ for the interval between −5.5 and −0.5 m, 0.9‰ between −0.5 and 6.25 m and 0.6‰ between 6.25 and 30 m (Figure 4). These intervals (“plateaus”) are separated by two main shifts, which occur between −1.5 and 1 m (from 1.4 to 0.4‰) and between 5.5 and 7 m (from 1.1 to 0.2‰). Less pronounced shifts to lighter values occur between −5 and −3 m and between 26.5 and 29 m. The most prominent shifts (of ~0.6‰) to heavier values are found between −3 and −1.5 m, 1 and 1.7 m and 7 and 8.2 m (Figure 4).

[32] High-frequency variations in δ¹⁸O show a distinct relation with lithology with lighter values in the reddish
layers (amplitude changes of ~0.5%; Figure 4). Close inspection of the $\delta^{18}$O record reveals additional high-frequency variability with a spacing of about twice that of the lithology-bound changes. Most of the prominent shifts described above are part of this high-frequency variability.

6.2.2. The $\delta^{13}$C Record

The $\delta^{13}$C record does not reveal a clear trend but significantly lighter values are reached in the top part of the record (Figure 4). Higher-frequency variations are less prominent than in the $\delta^{18}$O record. Nevertheless, lithology-bound changes are present with lighter values recorded in the beige layers. The expression of a cycle with approximately twice the thickness of a color cycle is more evident. The signal follows the same frequency changes as in $\delta^{18}$O, with shifts to lighter $\delta^{13}$C values coinciding with shifts to heavier $\delta^{18}$O values (Figure 4). This relation does not hold, however, in the uppermost part of the section, i.e., from ~27 m upward.

7. Age Model

7.1. Loulja-A

The Loulja-A section must be slightly younger than the Ain el Beida section of late Messinian age in view of the overall orientation of the bedding plane and its location approximately halfway between Ain el Beida and Salé (lower Pliocene). Moreover, the characteristic interval with 6 successive thin cycles (LA-10B-15B) is unknown from Ain el Beida. Paleobathymetric estimates of 300–500 m follow logically from the deeper estimates for the latter section. An age of around the M-P boundary is also suggested by the calcareous nannofossil biostratigraphy. The $D. quinqueramus$ LO, $C. larrymayeri$ FO and LO, and $T. rugosus$ FO are recorded in the same order across the boundary at ODP Site 926 in the equatorial Atlantic [Backman and Raffi, 1997], while the younger events are found in sediments of early Pliocene age at DSDP Site 502 in the North Atlantic and in southern Italy, albeit not always in the same order [Driever, 1988; Lourens et al., 1996; Raffi et al., 1998; E. De Kaenel, unpublished data, 2001]. The nannofossil events exclude any other correlation than with the Miocene-Pliocene boundary interval.

[35] The initial astronomical tuning of Loulja-A is based on the color cycles. Because of the close proximity of the sections in the Bou Regreg area and the identical expression of the cycles, it is assumed that the phase relations between the color cycles and the orbital parameters are the same as for Ain el Beida. Phase relations for the color cycles in the

Figure 4. Geochemical records (PC-1, benthic $\delta^{18}$O and $\delta^{13}$C) of the Loulja-A section. For key, see caption to Figure 2. Open squares in the isotope records represent data obtained from $Cibicides dutemplei$; open circles denote the $Cibicides pachyderma$ values. Correction of the $Cibicides$ values was done using a value equal to the average offset from the $Planulina$ values; these values are 0.214 ($\delta^{13}$O) and 0.324 ($\delta^{13}$C) for $C. dutemplei$ and 0.065 ($\delta^{18}$O) and 0.627 ($\delta^{13}$C) for $C. pachyderma$. 

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latter section were based on similarities in proxy signals with sapropels in the Mediterranean for which the phase relations are known [Lourens et al., 1996; Krijgsman et al., 2004]. This implies that midpoints of individual, Ca-poor (section 6.1), reddish layers correspond to precession minima and summer insolation maxima and that thicker and more prominent reddish layers correspond to high-amplitude precession minima and insolation maxima and, hence to eccentricity maxima [Krijgsman et al., 2004].

[36] We start the tuning from the characteristic interval of 6 thin cycles (LA-10B up to LA-15B). According to the phase relations and initial age constraints, these cycles correspond to the 400-kyr eccentricity minimum around 5.25 Ma (Figure 5). Tuning the individual sedimentary cycles to the precession and/or insolation target curve usually follows such a first-order tuning. Instead we preferred at this stage to introduce the PC1 record as an additional tool in the tuning exercise because of its distinct cyclic character. A detailed study of the Ain el Beida section revealed a very good fit between a component with similar element loadings (E. Van der Laan, unpublished data, 2002) and the astronomical target curve but this component could better be compared with ETP (a combined record of normalized Eccentricity, Tilt (obliquity) and (negative) Precession [Imbrie et al., 1984]) than with insolation because of an apparent nonlinear response to the insolation forcing. According to our phase relations, maxima in PC1 correspond to maxima in ETP (maxima in obliquity and eccentricity/minima in precession).

[37] The PC1 record of Loulja-A is in almost excellent agreement with ETP, allowing for a straightforward tuning of the section up to at least cycle LA-16 (Figure 5). The two groups of three distinct maxima in PC1 in the lower part of Loulja-A (LA-2-4; LA-7-9) correlate with two similar clusters of maxima in ETP between 5.58 and 5.38 Ma, which reflect two successive prominent 100-kyr eccentricity maxima. The two weaker maxima of cycles LA-5 and LA-6 correlate with the two low-amplitude maxima in ETP that correspond to the intervening 100-kyr eccentricity minimum around 5.48 Ma. The amplitude pattern in PC1 generally mimics that of ETP. The thick reddish layer LA-10B is partly explained by the composite character of this cycle in ETP. The correlative ETP maximum shows a “shoulder” at its base corresponding to an additional low-amplitude precession cycle. Similar shoulders (but in opposite direction) may explain the thickness of LA-1B and LA-9B. Following the tuning of cycles LA-1 to LA-10, the characteristic interval of thin color cycles (LA-10B to LA-15B) can be correlated straightforwardly to the interval with reduced amplitude variations in ETP and insolation that corresponds to the 400-kyr eccentricity minimum around 5.25 Ma.

[38] The tuning of cycles LA-17 to 21 proved more difficult. The preferred option is presented in Figure 5 and implies that all the color cycles correspond to single precession cycles, even though this does not explain the extra peaks in PC1 observed in some of the cycles. Following this option, LA-20B correlates with the precession minimum/insolation maximum at 5.14/5.15 Ma. The relatively thick reddish layer of cycle LA-17 does not reflect a high-amplitude precession minimum/insolation maximum as expected, but is related to the longer than average duration of the correlative precession-insolation cycle. Sedimentation rates, which are 5 cm/kyr on average in the interval with the 6 thin cycles LA-10B-15B, increase to ~13 cm/kyr in cycle LA-17 and to ~16 cm/kyr in cycle LA-20 (Figure 6). Other options imply that the reddish layers of one or all of cycles LA-18 to 21 contain an extra cycle that lacks sedimentary expression. Although double cycles are known from Ain el Beida, extra cycles are not developed in reddish layers in that section [Krijgsman et al., 2004].

[39] We favor the first option for several reasons. Firstly, the increase in sedimentation rate resulting from this option is expected because sedimentation rate is positively correlated with eccentricity and hence with precession amplitude both in Ain el Beida [Van der Laan et al., 2005] as well as in the lower half of Loulja-A (see Figure 4). Nevertheless, the increased sedimentation rate starts somewhat earlier than expected on the basis of the amplitude changes in ETP and insolation. However, the increase markedly coincides with the top of the 400-kyr carbonate maximum in the deep marine Trubi marls in southern Italy; this maximum corresponds to the 400-kyr eccentricity minimum around 5.25 Ma and consists of the first 6 precession-related cycles of the Trubi [Hilgen, 1991; Hilgen and Langereis, 1993]. Overall the higher sedimentation rates are in good agreement with sedimentation rates calculated for the lower part of Loulja-A and Loulja-B (see section 7.3) where cycles reach similar thicknesses at times of maximum eccentricity. The double cycle option does not result in higher sedimentation rates in the top part of Loulja-A, which is difficult to explain. Moreover, out-of-phase relationships between the 41-kyr component in Δ18O and obliquity start to develop if this option is used for tuning (not shown). Finally, the remaining cycles LA-22-27 should also represent double cycles since they reach similar thicknesses as cycles LA-18-21. However, as we demonstrate in section 7.3, it is clear from the tuning of Loulja-B that the extremely thick cycles LB-8-9 (which do represent double cycles) should then be reached in the top part of Loulja-A, which is clearly not the case. We therefore prefer the tuning option in which all color cycles are regarded as single precession cycles. The sampled time interval covered by the Loulja-A section thus ranges from ~5.59 to 5.12 Ma.

7.2. Spectral Analysis

[46] To further test the correctness of our preferred age model, we applied (cross) spectral analysis and band-pass filtering using the Analyseries software of Paillard et al. [1996] to determine the main periodicities in our proxy records and the phase relationships relative to the astronomical forcing. The PC1 spectrum reveals peaks at the main precession frequencies, which is not surprising because of the strong link with the color cycles and the tuning to a precession-dominated target curve (Figure 6). Additional variance near the main obliquity frequency is also apparent although a separate peak is lacking. In addition, power is concentrated around ~400 kyr which is probably related to the long-term eccentricity cycle. The Δ18O spectrum reveals a strong peak at the main obliquity frequency.
Figure 5. Calibration of the magnetostratigraphy of the Loulja-A and Loulja-B sections (for key, see caption to Figure 2) to the astronomical polarity timescale (ATNTS2004 of Lourens et al. [2004]) and tuning of color cycles and PC-1 to the 65°N summer insolation and ETP (normalized eccentricity, tilt and precession) time series of the La2004 solution [Laskar et al., 2004] with present-day values for dynamical ellipticity and tidal dissipation. Also shown are the precession and eccentricity curves of the same solution.
Figure 6. (a) Sedimentation rate curve versus lithology of Loulja-A according to our preferred age model. Dashed line shows sedimentation rate if cycles LA-17-20 represent double cycles. (b) Power spectra of PC-1, benthic δ¹⁸O and δ¹³C, and sedimentation rate, using the preferred age model to generate time series. Solid and dotted lines in each spectral panel represent spectral power after tuning to ETP and insolation, respectively. Horizontal bar denotes the band width (BW), which is 0.011 for PC-1, δ¹⁸O, and δ¹³C and 0.012 for sedimentation rate. Vertical bar denotes the 95% confidence interval (CI); for the sedimentation rate the 80% confidence interval is indicated.
Subsidiary peaks occur at the low-frequency end of the spectrum, at ~260 kyr and ~80 kyr. The $\delta^{13}$C spectrum reveals a concentration of power at the low-frequency end of the spectrum, while a shoulder is recorded at ~90 kyr, presumably related to short-term eccentricity. Additional peaks at the main obliquity and precession frequencies are present although their power is relatively low. Because the $\delta^{18}$O and $\delta^{13}$C time series were constructed by means of tuning PC-1 to ETP, we carried out an independent check on the correctness of our age model by filtering the 41-kyr components in $\delta^{18}$O, $\delta^{13}$C and PC-1 and comparing them with obliquity. The 41-kyr components generally show an in-phase relationship with obliquity and similar amplitude variations (Figure 7), although the 41-kyr components in PC1 and $\delta^{13}$C start to run out-of-phase with obliquity in the younger part of the record (from 5.52 Ma onward). For $\delta^{18}$O, the out-of-phase relationship stems from the fact that it does not covary inversely with $\delta^{13}$O in this part of the record, as mentioned before (see section 6.2.2). In addition, a clear obliquity-related signal is lacking in the interval between 5.48 and 5.36 Ma, where amplitude variations are low and the ~100-kyr signal is well developed (Figure 7). For PC1, the (near) out-of-phase relationship with obliquity in the younger (and oldest) part of the record stems from the high-amplitude precession-related variations in PC1 associated with the eccentricity maxima at ~5.15 and 5.55 Ma; these strong variations prevent the filter from picking up the obliquity-related signal.

Results of cross-spectral analysis reveal that the precession components in PC1, $\delta^{13}$C and $\delta^{18}$O vary in-phase with ETP (Figure 8). Again this is not surprising in view of the strong link of these components with lithology and the tuning (of PC1) to a precession-dominated target curve. The results also confirm the in-phase relation for the 41-kyr component in $\delta^{13}$O with respect to obliquity as inferred from the band-pass filtering. Note, however, that the latter in-phase relation changes to a lag of several kyr if the tidal dissipation in the La2004(1,1,0) solution is reduced to half its present-day value (not shown). This outcome is
comparable with the results from the slightly older Ain el Beida section [Van der Laan et al., 2005] and provides additional supportive evidence for our tuned age model of the Loulja-A section. Our age model provides astronomical ages, not only for the sedimentary cycles and bioevents, but also for the dominantly obliquity-controlled $\delta^{18}O$ stages (Tables 2 and 3).

7.3. Loulja-B

The calcareous plankton biostratigraphic data from Loulja-B indicate that this section is slightly younger than Loulja-A (thereby confirming the information from the field) and that the N-R polarity reversal must represent the Upper Thvera. Using the position of the Upper Thvera dated astronomically at 4.997 Ma [Lourens et al., 2004] as starting point, the tuning of Loulja-B is rather straightforward (Figure 5). The two very thick cycles LB-8 and LB-9 are interpreted as double cycles in which one cycle lacks sedimentary expression and correspond to the 400-kyr eccentricity minimum around 4.85 Ma; this minimum is characterized by strongly reduced amplitudes in precession and hence insolation. In this case the very weak summer insolation minima (precession maxima) at 4.87 and 4.83 Ma are not expressed in the sedimentary record as additional thin beige marls. This pattern strongly resembles that observed in the Trubi marl formation where cycles 21 and 22 represent double cycles as well [Hilgen, 1991; Lourens et al., 1996]. The very thick reddish layer of cycle LB-2 is the only uncertainty that remains in the tuning of Loulja-B. Similarly, this cycle may represent a double cycle.

7.4. Calcareous Plankton Events and Magnetic Reversals

With the tuning of the Loulja sections, astronomical ages are obtained not only for the sedimentary and geochemical cycles but also for the calcareous plankton events and magnetic polarity reversals (Table 3). The Discoaster
LO at 5.537 Ma and the Driever, LO is synchronous between Loulja and the C., 1994, 2001; (A) et al.: No deglaciation Mio-Pliocene boundary

D. perplexus Hodell et al. corresponds to the V AN DER LAAN ET AL.: NO DEGLACIATION MIO-PLIOCENE BOUNDARY

Triquetrorhabdulus quinqueramus PA3011<br>Table 2. Stratigraphic Position of Midpoints of Reddish and Beige Layers of Lithologic Cycles in the Loulja-A Section and Their Astronomical Ages After Tuning to the ETP and Summer Insolation Curves of the La2004(1,1) Solution

<table>
<thead>
<tr>
<th>Cycle</th>
<th>Stratigraphic Position, m</th>
<th>Age, Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>LA-2R</td>
<td>-2.70</td>
<td>5.561</td>
</tr>
<tr>
<td>LA-2B</td>
<td>-1.80</td>
<td>5.550</td>
</tr>
<tr>
<td>LA-3R</td>
<td>-0.65</td>
<td>5.539</td>
</tr>
<tr>
<td>LA-3B</td>
<td>0.40</td>
<td>5.530</td>
</tr>
<tr>
<td>LA-4R</td>
<td>1.15</td>
<td>5.519</td>
</tr>
<tr>
<td>LA-5R</td>
<td>3.65</td>
<td>5.497</td>
</tr>
<tr>
<td>LA-6R</td>
<td>4.80</td>
<td>5.470</td>
</tr>
<tr>
<td>LA-6B</td>
<td>5.30</td>
<td>5.459</td>
</tr>
<tr>
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</tr>
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<td>5.435</td>
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<td>9.20</td>
<td>5.416</td>
</tr>
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<td>LA-9R</td>
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</tr>
<tr>
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<td>12.90</td>
<td>5.355</td>
</tr>
<tr>
<td>LA-10B</td>
<td>13.30</td>
<td>5.342</td>
</tr>
<tr>
<td>LA-11R</td>
<td>13.90</td>
<td>5.331</td>
</tr>
<tr>
<td>LA-11B</td>
<td>14.40</td>
<td>5.322</td>
</tr>
<tr>
<td>LA-12R</td>
<td>15.10</td>
<td>5.313</td>
</tr>
<tr>
<td>LA-12B</td>
<td>15.60</td>
<td>5.301</td>
</tr>
<tr>
<td>LA-13R</td>
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<td>5.292</td>
</tr>
<tr>
<td>LA-13B</td>
<td>16.60</td>
<td>5.286</td>
</tr>
<tr>
<td>LA-14R</td>
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<td>5.275</td>
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<td>LA-14B</td>
<td>17.65</td>
<td>5.264</td>
</tr>
<tr>
<td>LA-15R</td>
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<td>5.254</td>
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<tr>
<td>LA-15B</td>
<td>18.70</td>
<td>5.249</td>
</tr>
<tr>
<td>LA-16R</td>
<td>19.65</td>
<td>5.237</td>
</tr>
<tr>
<td>LA-16B</td>
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<td>5.219</td>
</tr>
<tr>
<td>LA-17B</td>
<td>24.20</td>
<td>5.193</td>
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<tr>
<td>LA-18B</td>
<td>26.50</td>
<td>5.173</td>
</tr>
<tr>
<td>LA-19B</td>
<td>29.10</td>
<td>5.149</td>
</tr>
<tr>
<td>LA-20B</td>
<td>32.30</td>
<td>5.131</td>
</tr>
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</table>

Astronomical Ages After Tuning to the ETP and Summer Insolation Curves of the La2004(1,1) Solution

PA3011

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8. Discussion
8.1. Isotope Stratigraphy and Chronology

The planktonic foraminifera do not show clear-cut primary events, but the neogloboquadrinid coiling direction reveals two sinistral influxes of very short duration. Two similar influxes of sinistral neogloboquadrisnids have also been found in the Mediterranean where they occur in the homogenous mnl of basic cycles 1 and 2 in the Trubi Formation [Di Stefano et al., 1996; Lourens et al., 1996]. The independent tuning of the Loulja color cycles indicates that the first influx corresponds exactly in time with the first influx in the Mediterranean but that the second influx occurs one precession-related cycle later at Loulja.

The astronomical calibration implies that Loulja-A cycles 1–21 cover the time interval from 5.59 to 5.12 Ma. Therefore the base of the normal polarity chronozone recorded between 20.35 and 20.7 m in cycle LA-16 should correspond to the base of the Thvera subchron (C3n.4n (o)); the astronomical age of 5.222 ± 0.005 Ma for the reversal at Loulja is very close to the astrochronologic age of 5.235 Ma [Lourens et al., 1996, 2004]. In addition, the reversal is neatly located five precession cycles (~10 kyr) above the Mio-Pliocene boundary, dated astronomically at 5.332 Ma [Lourens et al., 1996; Van Couweling et al., 2000] and pinpointed at 13.90 m in the middle of LA-11R. The N > R transition below the 11 m level in LB-3g corresponds to the Upper Thvera, as previously assumed. Its position is consistent with the Mediterranean, indicating that the very thick cycles LB-8 and LB-9 are indeed the equivalent of the double cycles 21 and 22 in the Trubi Formation. Note however that the position of this reversal was used during the tuning procedure.

The three cosmopolitan nannofossil events recorded in the lower part of the Loulja-B section, namely the subbottom of Reticulofenestra pseudoumbilicus (>8 μm), the D. perplexus subtop and the Pseudoemiliania lacunosa ovata FO, were also recorded by Driever [1988] in the Mediterranean Singa and Rossello composite sections dated astronomically by Lourens et al. [1996]. Our astronomical ages for these events are in good agreement with Driever [1988] and Lourens et al. [1996]. The difference of one cycle for the D. perplexus subtop is due to a difference in defining the n1 event of Driever [1988], which marks the transition from predominantly Dictyococcites perplexus to predominantly R. pseudoumbilicus (>8 μm), rather than a problem of the astronomical tuning. The rare occurrence of Pseudoemiliania lacunosa ovata at the beginning of its range explains the slightly different age obtained for this event in the present study.

The Loulja-A δ18O record reveals a much stronger obliquity-related signal than PC1, which mainly tracks the color cycles observed in the field. Because of the additional influence of precession, it is not always easy to recognize...
the obliquity-related signal in $\delta^{18}O$. The latter signal was illustrated by extracting the 41-kyr component in $\delta^{18}O$ and showing it as an overlay on the $\delta^{18}O$ time series; comparison with obliquity reveals an almost perfect in-phase relationship (Figure 7).

Like at Ain el Beida, the strong 41-kyr $\delta^{18}O$ cycle at Louljá-A is interpreted as to reflect a dominant obliquity-controlled glacial cyclicity, which has been recognized in open ocean benthic isotope records of latest Miocene to early Pleistocene age [e.g., Ruddiman et al., 1986; Raymo et al., 1989; Tiedemann et al., 1994; Shackleton et al., 1995b]. This glacial record reveals a characteristic pattern of distinct and less distinct glacial cycles. The more prominent peaks in $\delta^{18}O$ have been labeled to facilitate communication and global correlation of the oxygen isotope stages [Tiedemann et al., 1994; Shackleton et al., 1995b]. For the latest Miocene to early Pliocene, i.e., the time span covered by the Louljá-A section, the codification scheme of Shackleton et al.[1995b] based on ODP Site 846 in the eastern equatorial Pacific is used. This scheme links isotope stages to magnetochrons whereby, similar to other schemes, even numbers indicate cold stages. The initial tuned age model for the isotope record of Site 846 [Shackleton et al., 1995a] was revised later on by Shackleton et al. [1999].

### Table 3. Tuned Ages of Magnetic Reversals, Calcareous Plankton Events, and Marine Isotope Stages in the Louljá Sections and Comparison With Ages From Other Studies

<table>
<thead>
<tr>
<th>Cycle</th>
<th>Position, m</th>
<th>Age, Ma</th>
<th>Other Age,a Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Magnetostratigraphy</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C3n.4n (old)</td>
<td>LA-16B</td>
<td>20.35–20.70</td>
<td>5.222 ± 0.003</td>
</tr>
<tr>
<td>C3n.4n (young)</td>
<td>LB-3</td>
<td>9.8–11.6</td>
<td>4.994 ± 0.007</td>
</tr>
<tr>
<td><strong>Biostratigraphy</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Interval with common Globorotalia menardii</td>
<td>LA-2B</td>
<td>–1.95 to –1.55</td>
<td>5.550 ± 0.002</td>
</tr>
<tr>
<td>Interval with common Globorotalia menardii</td>
<td>LA-4B</td>
<td>2.10–2.20</td>
<td>5.511 ± 0.001</td>
</tr>
<tr>
<td>Interval with common Globorotalia menardii</td>
<td>LA-11B</td>
<td>14.55–14.75</td>
<td>5.319 ± 0.001</td>
</tr>
<tr>
<td>Influx sinistral Neogloboquadrinids</td>
<td>LA-13A</td>
<td>16.15–16.25</td>
<td>5.291 ± 0.001</td>
</tr>
<tr>
<td>Influx sinistral Neogloboquadrinids</td>
<td>LA-13B</td>
<td>16.65–16.75</td>
<td>5.285 ± 0.001</td>
</tr>
<tr>
<td>L.O. Discoaster quinqueramus</td>
<td>LA-3R</td>
<td>–0.7 to –0.18</td>
<td>5.537 ± 0.002</td>
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<td>F.O.D. Ceratolithus larrymayeri</td>
<td>LA-12B</td>
<td>15.50–15.55</td>
<td>5.303 ± 0.001</td>
</tr>
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<td>L.O.D. Ceratolithus larrymayeri</td>
<td>LA-14B</td>
<td>17.7–17.9</td>
<td>5.262 ± 0.002</td>
</tr>
<tr>
<td>L.O. Triquetrorhabdulus rugosus</td>
<td>LA-15R</td>
<td>18.2–18.3</td>
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<td>F.O. Amaurolithus tricorniculatus</td>
<td>LA-20R</td>
<td>28.0–30.3</td>
<td>5.151 ± 0.009</td>
</tr>
<tr>
<td>F.O. Amaurolithus bizzarus</td>
<td>LB-2R</td>
<td>2.25–5.0</td>
<td>5.033 ± 0.004</td>
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<tr>
<td>subtop Reticulofenestra pseudoumbilicus (&gt;8 μm)</td>
<td>LB-5B</td>
<td>17.8–19.4</td>
<td>4.950 ± 0.005</td>
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<tr>
<td>F.O. Pseudoemiliania lacunosa ovata</td>
<td>LB-6B</td>
<td>21.7–22.5</td>
<td>4.929 ± 0.003</td>
</tr>
<tr>
<td>F.O. Helicosphaera sellii</td>
<td>LB-6B</td>
<td>21.7–22.5</td>
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</tr>
<tr>
<td>F.R.O. Helicosphaera sellii</td>
<td>LB-7B</td>
<td>25.0–27.0</td>
<td>4.906 ± 0.008</td>
</tr>
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**Marine Isotope Stages**

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<th>MISb</th>
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<tr>
<td>TG13</td>
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<td>5.560</td>
</tr>
<tr>
<td>TG12</td>
<td>–1.55</td>
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<td>TG11</td>
<td>1.10</td>
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</tr>
<tr>
<td>TG10.2</td>
<td>2.60</td>
<td>5.505</td>
</tr>
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<td>TG10.0</td>
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<td>5.458</td>
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<td>TG9</td>
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<td>5.445</td>
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<td>5.433–5.414</td>
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<td>TG8.0</td>
<td>11.60</td>
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<td>TG6.2</td>
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<td>TG5</td>
<td>17.35</td>
<td>5.276</td>
</tr>
<tr>
<td>TG4.2</td>
<td>18.05</td>
<td>5.259</td>
</tr>
<tr>
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<td>5.248–5.230</td>
</tr>
<tr>
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<td>5.217</td>
</tr>
<tr>
<td>TG3</td>
<td>23.10–25.70</td>
<td>5.202–5.180</td>
</tr>
<tr>
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<td>5.173</td>
</tr>
<tr>
<td>TG1</td>
<td>28.30–31.40</td>
<td>5.151–5.136</td>
</tr>
</tbody>
</table>

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*a* Numbers in parentheses refer to 1, Lourens et al. [2004]; 2, Krijgsman et al. [2004]; 3, Lourens et al. [1996]; 4, Backman and Raffi [1997]; 5, Raffi et al. [1998]; 6, Driever [1988]; 7, Rico et al. [1997].

*b* Marine isotope stages follow Shackleton et al. [1995b]; (sub) stages in italics and stage [x] are introduced by us and indicated in Figures 10 and 11.
At Loulja-A, the correct identification of isotope stages is crucial if one aims to determine whether the basal Pliocene flooding of the Mediterranean is related to the (peak) deglaciation and associated sea level rise leading to TG5. The straightforward and unambiguous tuning of (the lower part of) Loulja-A reveals a stratigraphic overlap with Ain el Beida, indicating that the two prominent and characteristic maxima in $\delta^{18}O$ at $-5$ and $-1.5$ m correspond to TG14 and 12 (Figure 9). The identification of the younger stages depends on the detailed correlation to the $\delta^{18}O$ record of Site 846 in which they were originally defined [Shackleton et al., 1995b]. The stepwise shift to lighter isotope values between 5.55 Ma (TG12) and 5.45 Ma at Loulja is easily recognized at Site 846, indicating that the prominent $\delta^{18}O$ minimum at 5.45 Ma represents peak interglacial TG9. The $\delta^{18}O$ pattern in this interval is remarkably similar between both records showing additional precession-related variations that would allow a (further) subdivision of several of the stages (e.g., TG10). The Loulja-A record further confirms the presence of an extra obliquity-related cycle between TG11 and 9.

Figure 9. Comparison between the Loulja-A and ODP Site 846 $\delta^{18}O$ records in the time domain. For Site 846 the manually revised age model of Shackleton et al. [1999] was used to construct the proxy time series. The 41-kyr components are shown as overlays and were extracted using Gaussian band-pass filters with central frequencies of 0.0245 and a band width of 0.012 cycles/kyr. Indicated TG stages for Site 846 are from Shackleton et al. [1995b].

[50] At Loulja-A, the correct identification of isotope stages is crucial if one aims to determine whether the basal Pliocene flooding of the Mediterranean is related to the (peak) deglaciation and associated sea level rise leading to TG5. The straightforward and unambiguous tuning of (the lower part of) Loulja-A reveals a stratigraphic overlap with Ain el Beida, indicating that the two prominent and characteristic maxima in $\delta^{18}O$ at $-5$ and $-1.5$ m correspond to TG14 and 12 (Figure 9). The identification of the younger stages depends on the detailed correlation to the $\delta^{18}O$ record of Site 846 in which they were originally defined [Shackleton et al., 1995b]. The stepwise shift to lighter isotope values between 5.55 Ma (TG12) and 5.45 Ma at Loulja is easily recognized at Site 846, indicating that the prominent $\delta^{18}O$ minimum at 5.45 Ma represents peak interglacial TG9. The $\delta^{18}O$ pattern in this interval is remarkably similar between both records showing additional precession-related variations that would allow a (further) subdivision of several of the stages (e.g., TG10). The Loulja-A record further confirms the presence of an extra obliquity-related cycle between TG11 and 9.

[51] The isotope stage identification is difficult from TG9 onward due to the relatively low amplitude and therefore less characteristic pattern in $\delta^{18}O$ between 5.4 and $\sim5.2$ Ma at ODP Site 846 [Shackleton et al., 1995b, 1999]. Comparison of the Loulja-A and ODP Site 846 $\delta^{18}O$ time series, using the revised tuning of Shackleton et al. [1999], shows that TG9 to TG3 occur in both records at the same time and
are in-phase with obliquity, but that the filtered 41-kyr component starts to run out-of-phase with obliquity between TG3 and T7 at Site 846 (Figure 9). This would imply that the $\delta^{18}O$ minimum in the Loulja-A record at 5.315 Ma indeed corresponds to TG5 at Site 846.

[52] This identification becomes less certain if one compares the carbon and oxygen isotope records of Site 846 in the depth and time domain. In addition, the isotope records from ODP Site 982 in the North Atlantic [Hodell et al., 2001] and from Salé Briqueterie [Hodell et al., 1994] located near Loulja were included in the comparison because ages for $\delta^{13}C$ stages at Site 846 and for those in the redated Salé record [Hodell et al., 2001] deviate significantly from our tuned ages. The comparison for $\delta^{18}O$ and $\delta^{13}C$ in the depth and time domain is shown in Figure 10 ($\delta^{18}O$ in Figures 10a and 10b; $\delta^{13}C$ in Figures 10c and 10d). For Site 846, the depth-time conversion results in an increase in sedimentation rate between TG7 and TG5 if the (manual) revision of the initial tuned age model by Shackleton et al. [1999] is adopted (Figure 9). The TG7-5 interval is significantly thicker as that observed between regular obliquity-controlled cycles while these stages have been tuned to successive obliquity maxima. In fact, the Site 846 $\delta^{13}C$ record suggests that an additional obliquity-controlled cycle (labeled [x] in Figure 10c) is present between TG7 and TG5. This characteristic interval reveals the highest $\delta^{13}C$ values throughout the critical interval across the M-P boundary and is preceded by a long-term increase leading to TG7. It can relatively easily be recognized in the other $\delta^{13}C$ records, showing a very similar character at Loulja and ODP Site 846 with 3 distinct $\delta^{13}C$ maxima. The first and last of these maxima correspond to stages TG7 and 5 if we follow the $\delta^{18}O$ stage numbering of Shackleton et al. [1995b] and the expression of these stages in the Site 846 $\delta^{13}C$ record. As a consequence, the extra (obliquity-controlled) stage is also recognized in the Site 846 and Loulja $\delta^{13}C$ depth records. The extra cycle would have the immediate consequence that the M-P boundary does not coincide with TG5 but with the unnamed extra cycle between TG5 and TG7, if we assume that the identification and tuning of TG7 is correct.

[55] Using characteristic patterns in especially $\delta^{13}C$, the correlation between Loulja-A, Salé and ODP Sites 846 and 982 can be extended up to TG2 (Figures 10 and 11). These patterns include the marked shift to lighter values between TG3 and TG2 and the “glacial” $\delta^{13}C$ signature of TG4 and TG6. Ideally one would prefer to check the implications of the revised tuning up to the prominent glacial stages Si4 and Si6. However, this exercise is less straightforward and falls outside the scope of the present study, which aimed to unravel the influence of glacio-eustatic sea level change on the Pliocene flooding of the Mediterranean and the beginning of the Upper Evaporites.

[54] Finally the correlations between Loulja-A and Salé on the basis of the stable isotope patterns are in agreement with the occurrence of a peak interval in magnetic susceptibility observed at both localities (Figure 12). However, the relatively low sample resolution at Salé hampers a more detailed correlation of both susceptibility records.

8.2. Absolute Isotope Values

[55] Although the isotope records discussed above correlate in considerable detail, they differ in absolute values. The absolute values of the oxygen isotopes are very similar for the Loulja and Salé sections, but are different for Sites 846 and 982. This can be explained by a shallower depth and higher bottom water temperatures at the Moroccan sites. Nevertheless, the overprint of the dominantly obliquity controlled global ice volume signal is clearly reflected in all records and was used to correlate them in detail. The precession-related variations at Loulja most likely reflect a combination of a local temperature and/or salinity signal associated with the sedimentary color cycles and a global ice volume signal. Indications for the presence of a precession-related component in global ice volume change during the latest Miocene previously came from comparing the slightly older record from Ain el Beida with records from the open ocean [Van der Laan et al., 2005].

[56] The absolute carbon isotope values are very similar for Loulja, Salé, and ODP Site 846 but the values at Site 846 are considerably lower. This difference can best be explained by $\delta^{13}C$ aging of Pacific deep water [see, e.g., Ruddiman, 2001]. However, the overall pattern of all records is quite similar suggesting that it is controlled by whole-ocean processes (e.g., terrestrial-marine carbon transfer).

8.3. Miocene-Pliocene Boundary

[57] Irrespective whether the low-amplitude $\delta^{18}O$ minimum at 5.315 Ma in the Loulja-A record corresponds to TG5, it is clear that the M-P boundary coincides with the very beginning (onset) of a minor shift to lighter values that, as far as the obliquity component is concerned, leads to the low-amplitude $\delta^{18}O$ minimum at 5.315 Ma. This transition (“deglaciation”) is interrupted by a precession-controlled shift to heavier values in $\delta^{18}O$ but as yet it is unclear whether this precession component is also present in the $\delta^{18}O$ records from the open ocean. Such a presence may...
Figure 10
Figure 10. (continued)
suggest a precession component in the glacial cyclicity, i.e., similar to the precession signal between TG14 and 9 and for instance between TG32 and 28 (i.e., TG30) and between C3An.18O.14 (codification of Hodell et al. [1994]) and C3An.18O.10 (i.e., 12) at Ain el Beida [Van der Laan et al., 2005]. On the other hand, at Loulja-A it can also easily be related to the dominantly precession-controlled regional climate oscillations that underlie the color and PC1 cyclicity.

[58] Close inspection of the isotope record of Site 846 in fact indicates that the M-P boundary coincides with an extra obliquity-controlled and very weak interglacial between TG7 and TG5 and that TG5 represents an interglacial stage some 50 kyr younger than the M-P boundary (see section 8.1 and Figures 10 and 11). Comparison with our record further suggests that the actual Site 846 peak $\delta^{18}O$ values reached in TG5 may represent outliers because such exceptionally light values are neither found in our record nor in other records spanning the boundary [Hodell et al., 1994, 2001].

[59] All this indicates that glacio-eustatic sea level rise (deglaciation) played only a minor role, if at all, in the basal Pliocene flooding of the Mediterranean. Probably more important factors were extensional tectonics and/or ongoing headward erosion of an initially restricted connection between the Atlantic and Mediterranean. The latter situation may already have played a role during deposition of the Upper Evaporites in the Mediterranean with Atlantic waters intermittently entering the Mediterranean, thereby increasing seawater salinities.

[60] The M-P boundary is formally defined at the base of the deep marine Trubi marls overlying the Upper Evaporites and Arenazzolo Formation of the Messinian at Eraclea Minoa [Van Couvering et al., 2000]. This level marks the onset of open marine conditions in the Mediterranean following the Messinian salinity crisis [Hsü et al., 1973; Cita, 1975; Hilgen and Langereis, 1993; Iaccarino et al., 1999]. This level was selected mainly for historical reasons but the possibly unconformable nature of the formation boundary in combination with the abrupt transition from nonmarine to fully marine conditions makes it in principle less suitable. Suc et al. [1997], recognizing the problem of the possible unconformable character of the contact, proposed to define the M-P boundary near Sale, some 5 km northwest of Ain el Beida, at the stratigraphic level that corresponds with the base of oxygen isotope stage TG5. The isotope shift was supposed to correspond with the base of the Trubi reflecting the marked deglaciation held responsible for the Pliocene flooding of the Mediterranean at that time. The proposal was considered less favorable because the level was only indicated in the Sale drill core but not in any existing outcrop.

[61] Our results show that the M-P boundary is positioned at 13.90 m in the middle of LA_R-11 in section Loulja-A in an open marine succession located just outside the Mediterranean. Hence it might be preferable to designate Loulja-A as an auxiliary stratotype section for the M-P boundary. Our results also indicate however that the boundary does not coincide with TG5.

**8.4. Upper Evaporites**

[62] The stepwise shift to lighter values between TG12 and 9 marks a more dramatic deglaciation than TG5 [e.g., Shackleton et al., 1995b]. An important implication of the revised age model of Shackleton et al. [1999] is that the onset of the Upper Evaporites (UE) following the main desiccation phase of the Mediterranean is linked to the glacio-eustatic sea level rise associated with the deglaciation manifested in particular by the TG12 to 11 transition. The number of 7/8 sedimentary cycles observed in the UE and

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**Figure 11.** Comparison between the (a) $\delta^{18}O$ and (b) $\delta^{13}C$ time series of Loulja-A, Salé, and ODP Sites 846 and 982 after adopting revised age models based on control points (indicated by crosses) from Loulja-A. (c) Resulting power spectra for $\delta^{18}O$; note the relatively strong power in the precession (23–19 kyr) band. In each spectral panel the horizontal bar denotes the band width (BW), which is 0.0093 for Salé, 0.0101 for Site 982, and 0.0097 for Site 846. Vertical bar denotes the 95% confidence interval (CI).
correlative units throughout the Mediterranean [Vai, 1997; Krijgsman et al., 2001; Fortuin and Krijgsman, 2003] is consistent with such a scenario in case these cycles are precession-controlled (Figure 13). Alternative options start from an obliquity control for the UE cyclicity [Vai, 1997]; this interpretation is inconsistent with linking the beginning of the UE to the TG12 to 9 deglaciation. It further is not in agreement with a precession control for the evaporite cycles of the Lower Evaporites (LE) although the distinction between a precession or obliquity control on the UE cyclicity becomes problematical in view of the hiatus of ~100 kyr inferred between the LE and UE according to the precession scenario [Krijgsman et al., 1999, 2001].

There are indications in the literature that the Mediterranean was not fully isolated from the world ocean during deposition of the UE but that marine incursions may have occurred. The gypsum cycles of the UE in the Eraclea Minoa section (Caltanissetta Basin, Sicily) display a facies evolution from fine-grained laminar gypsum at the base of each cycle to selenites at the top, accompanied by increasing strontium values indicating increasing salinity conditions [Rosell et al., 1998]. Partly backed up by these data, Schreiber [1997] claimed a distinctly marine origin of the gypsum in this section alternating with nonmarine beds and concludes that a connection with the open ocean existed throughout the MSC. Also, Londeix [2004] interpreted the intercalated clays of Eraclea Minoa with dinocyst assemblages to be indicative of a shallow marine environment with occasionally hypersaline conditions and river influxes. However, according to Taberner et al. [2004], previous interpretations of the geochemical and biological markers from the Eraclea Minoa UE, supposedly indicative of a marine contribution need reconsideration. Recycling of older Messinian evaporites may offer an alternative explanation for the observed isotopic composition of the sulphates. Nevertheless, Flecker et al. [2002] argued that neither the observed Sr isotopic offsets from ocean values nor the high salinities during evaporite precipitation require total isolation of the Mediterranean from the world ocean.

Also, Clauzon et al. [1996] considered marginal basins (including Eraclea Minoa) in the Mediterranean to have received Atlantic waters during deposition of the LE and UE, although their age model differs from the age model we prefer. From the uppermost Messinian of the Nijar Basin in SE Spain, Fortuin and Krijgsman [2003] reported marine microfaunas and suggested that short-lived connections with the Atlantic Ocean could have existed during periods of high sea level although reworking could not be excluded. Aguirre and Sanchez-Almazo [2004] are much more explicit by stating that the planktonic foraminiferal faunas from the Messinian postevaporitic beds in the Nijar Basin are not reworked but represent in situ faunas. The latter would imply that open marine conditions already existed in the western Mediterranean area during the latest Messinian. Nevertheless, the arguments put forward by Aguirre and Sanchez-Almazo [2004] concerning the sometimes excellent preservation and the absence of reworked older faunas do by no means exclude reworking. Even on Cyprus, Rouchy et al. [2001] claim to have found authentic marine faunas in the local equivalent of the UE/Lago Mare, indicative of the persistence of marine influences. Müller [1990] mentioned that assemblages of low diversity and smaller sized calcareous nannofossils and planktonic foraminifera are found in the uppermost Messinian of the
Tyrrhenian Sea, while Castradori [1998] observed anomalous nanofossil assemblages in the same interval in the Alboran, Balearic and Levantine basins. From the latter basin and from the Mediterranean Ridge south of Crete, Spezzaferri et al. [1998] ascribed the presence of beds with planktonic foraminifera to an Atlantic origin, a conclusion that was shared, although with more caution, by Iaccarino and Bossio [1999] for sediments of the uppermost Mesianian in the Balearic Basin. Iaccarino et al. [1999] stated that periodic spillovers into the (western) Mediterranean occurred at the end of the Mesianian, related to eustatic sea level changes. [65] The latter would favor an obliquity control because of the dominant obliquity forcing of glacial cyclicity and hence glacio-eustatic sea level change during the latest Miocene and Pliocene. Nevertheless, it is evident that the benthic isotope records reveal a well-defined precession signal in addition to obliquity in this interval, suggesting that precession exerted an additional control on glacial cyclicity and sea level change in the interval between TG12 and 5, i.e., the time that the UE were formed according to our preferred precession scenario. However, the precession control may also stem from precession-induced changes in regional climate. Such changes are also held responsible for the formation of gypsum cycles in the LE, i.e., between 5.96 and 5.59 Ma when obliquity-induced glacial cycles and associated sea level changes were operative as well [Krijgsman et al., 1999]. [66] Finally it is remarkable that the Upper Evaporites according to the preferred precession scenario correspond

Figure 13. Cyclostratigraphic correlations of the Loula-B and Loula-B sections (for key, see caption to Figure 2) to the Upper Evaporites and Trubi marls in the Eraclea Minoa section on Sicily, Italy (note scale change at the Miocene-Pliocene boundary). Eraclea Minoa tuning is partly based on the work of Lourens et al. [1996]. Correlations of the PC-1 and $\delta^{18}O$ records between the basal part of Loula-A and the uppermost part of Ain el Beida [Krijgsman et al., 2004; Van der Laan et al., 2005; Van der Laan, unpublished data, 2003] are also shown. Isotope stages are numbered following the codification scheme of Shackleton et al. [1995b]; see caption to Figure 9 for further explanation.
exactly with the interval marked by peak magnetic susceptibility values in the Loulja-A section and the Salé drill core (Figure 12). In addition this peak interval corresponds to the 400-kyr eccentricity maximum around 5.45 Ma. The latter might explain our peak interval because magnetic susceptibility maxima are reached in reddish layers (Figure 2) and the most prominent reddish layers are associated with enhanced precession amplitudes related to maximum eccentricity. Such an explanation does hold for the elemental data of PC1, which clearly mimic the eccentricity modulation of the precession amplitude (see section 7.1 and Figure 5). However, in contrast to PC1, magnetic susceptibility values remain low in sections Loulja-A and Loulja-B, i.e., during the next 400-kyr eccentricity maximum. Furthermore it does not explain why the marked decrease in magnetic susceptibility (Figure 2) does not follow the precession/eccentricity pattern but coincides with the Mio-
docene-Pliocene boundary and thus with the top of the Upper Evaporites and the Pliocene flooding of the Mediterranean. At present, a sound explanation for this apparent linkage between the magnetic susceptibility peak interval and the Upper Evaporites is lacking.

8.5. Deglaciation Interval

[67] Also the biostratigraphic record of *G. miotumida/G. menardii* has some bearing on (the termination of) the late Messinian interval marked by the occurrence of distinct glacials. The warm water species *Globorotalia menardii* and *Neogloboquadrina numerosa* represent subtropical faunal elements which occur together with species typical of the eastern boundary current fauna association. Remarkably enough, the short interval with *G. menardii* corresponds exactly with the major shift in $\delta^{13}C$ to lighter values between oxygen isotope stages TG12 and 11 which marks the end of the characteristic interval with prominent obliquity controlled glacial cycles of the late Messinian. The short distribution range of this warm water species is consistent with the deglaciation step inferred from the isotope data. This deglaciation step coincides with the increase in eccentricity following the distinct eccentricity minimum around 5.6 Ma. Stage TG10 corresponds with the next (100 kyr) eccentricity minimum around 5.48 Ma and the (second) deglaciation step from TG10 to 9 with the subsequent increase in eccentricity just prior to 5.45 Ma. This coincidence suggests that eccentricity may have played a role in terminating the series of prominent glacial cycles in the late Messinian.

8.6. Glacial Cycles and Regional Climate Oscillations

[68] At Loulja-A, the PC1 and sedimentary cycle records show a dominant precession-related signal, suggesting a regional low-latitude climate origin for these variations [Van der Laan et al., 2005]. The 23-kyr component in $\delta^{13}C$ is rather weak and is only present in intervals when enhanced precession amplitudes are reached at times of eccentricity maxima. This 23-kyr signal is essentially in phase with precession, with minima in $\delta^{13}C$ corresponding to maxima in precession (see also Figure 8). The 23-kyr signal in $\delta^{18}O$ may be explained by enhanced surface water productivity, but a planktonic isotope record is needed to support this interpretation. The enhanced productivity might be a con-
sequence of better vertical mixing of the water column due to colder and hence denser surface waters during summer at times of precession maxima. In addition, precession-induced changes in wind stress and hence upwelling may have played a role. Alternatively, it may signal variations in the storage of carbon between the terrestrial and marine domain on precession timescales, although obliquity and eccentricity related variations in $\delta^{13}C$ seem to be more important here.

[69] Cross-spectral analysis in addition revealed a weak 41-kyr component in PC1 that is significantly coherent and essentially in phase with obliquity. This obliquity signal, which was also observed in the older Oued Akrech section (F. J. Hilgen, unpublished data, 2000), most likely reflects a direct response of regional climate to obliquity as suggested by Mediterranean sapropel patterns [Hilgen et al., 2000]. Contrary to expectation, no significant lag is observed between the obliquity component in $\delta^{13}C$ and PC1, despite the fact that the 41-kyr component in $\delta^{18}O$ is much stronger than in PC1 and supposedly reflects glacial cyclicity in addition to regional climate change. As a consequence, no distinction can be made between an early response group and a (glacial controlled) late response group reflecting ice volume and related ice sheet response [cf. Imbrie et al., 1992]. Similarly, the identical and in-phase relation between the 23-kyr components in PC1 and $\delta^{13}C$, and precession does not rule out a precession-induced ice volume signal in $\delta^{18}O$. The latter is actually suggested by the precession signal found in benthic $\delta^{18}O$ records from the open ocean (Figure 11c).

[70] Although a separate 41-kyr peak is lacking in the $\delta^{13}C$ spectrum, an inverse covariance with $\delta^{18}O$ is clearly recognized in the interval older than ~5.3 Ma. This inverse correlation is confirmed by the in-phase relationship between the 41-kyr $\delta^{13}C$ filter and obliquity up to at least 5.22 Ma (see section 7.4) and is interpreted as to mainly reflect glacial-interglacial cycles: during glacials (enriched $\delta^{18}O$ values), overall lighter $\delta^{13}C$ values are recorded in the marine realm due to increased storage of $^{13}$C-depleted terrestrial-derived carbon in oceanic reservoirs [Shackleton, 1977; Ruddiman, 2001]. Alternatively, enhanced productivity linked to a strengthening in wind regimes inferred for glacial periods, causing better vertical mixing through the water column, could have caused the lighter $\delta^{13}C$ values. However, after TG9 (~5.45 Ma), obliquity-induced changes in $\delta^{13}C$ become less distinct and longer-term variations with a period of ~100 kyr start to dominate. The ~100-kyr band-pass filter reveals an in-phase relationship with short-term eccentricity whereby $\delta^{13}C$ minima coincide with eccentricity maxima (see Figure 8). Only the interval of relatively depleted $\delta^{13}C$ values between 5.23 and 5.18 Ma causes the ~100-kyr $\delta^{13}C$ filter to slightly run out-of-phase with eccentricity.

9. Conclusions

[71] The Loulja-A section contains a well-tuned continuous open marine succession straddling the M-P boundary in the Bou Regreg area that can be correlated cyclostratigraphically in detail to the Mediterranean. In addition, the section provides a detailed history of glacial cyclicity
through its high-resolution and high-quality benthic oxygen isotope record.

[72] The M-P boundary does not coincide with a major deglaciation as previously assumed but with the beginning of a relatively minor shift to lighter values in $\delta^{18}O$. This shift does not coincide with marine isotope stage TG5 but actually corresponds to an extra (weak) obliquity-controlled cycle in between TG7 and TG5. On the other hand, the onset of the Upper Evaporites in the Mediterranean marked by hyposaline conditions coincides with the major deglaciation step between TG12 and TG11, suggesting that the associated sea level rise might at least be partly responsible for the onset of possibly (marginally) marine conditions following the main desiccation phase.

[73] The section is perfectly suitable to act as auxiliary boundary stratotype for the M-P boundary as formally defined at the base of the deep marine Trubi marls in the Eraclea Minoa section on Sicily, i.e., at the level that marks the definitive opening of the Atlantic-Mediterranean connection and the basal Pliocene flooding of the Mediterranean following the Messinian salinity crisis.

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