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Correlation of late Miocene to early Pliocene sequences between the Mediterranean and North Atlantic

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Abstract. Ocean Drilling Program (ODP) Site 982 in the North Atlantic contains a complete latest Miocene to early Pliocene section that was tuned to the astronomical timescale by correlating the record of gamma ray attenuation (GRA) bulk density to summer insolation at 65°N and the benthic $\delta^{18}O$ signal to orbital obliquity for the interval from 4.6 to 7.5 Ma. The astronomical tuning of the Site 982 record permits a direct bed-to-bed correlation to the cyclostratigraphy of Messinian sections in the Mediterranean [Krijgsman et al., 1999a, 2001]. The benthic $\delta^{18}O$ signal at Site 982 records a latest Miocene glacial period that lasted from ~6.26 to 5.50 Ma and consisted of 18 glacial-to-interglacial oscillations that were controlled by the 41-kyr cycle of obliquity. Although the intensification of glaciation at 6.26 Ma may have contributed to the restriction of the Mediterranean, it preceded the depositional onset of the lower evaporite unit at 5.96 Ma by some 300 kyr. The transition from Stage TG12 to TG11 at 5.5 Ma marks the end of the latest Miocene glacial period and precedes the Miocene/Pliocene boundary by 170 kyr. Although benthic $\delta^{18}O$ values are relatively low and $\delta^{18}O$ of bulk carbonate reaches a minimum at the Miocene/Pliocene boundary at 5.33 Ma, there is no single "event" that would indicate deglaciation and sea level rise as the cause of the reflooding of the Mediterranean. We conclude that glacioeustatic changes alone were not responsible for either the start or end of evaporite deposition during the Messinian, suggesting that tectonic or local climate changes in the Mediterranean region were the dominant cause(s).

1. Introduction

The role of glacioeustatic and tectonic changes in the isolation, desiccation, and reflooding of the Mediterranean during the Messinian has long been debated (for a review, see [Kastens 1992]). The marine oxygen isotope record provides a proxy for late Miocene temperature and glacioeustatic changes that can be compared with the timing of events during the Messinian Salinity Crisis (MSC). Our ability to infer causal relationships between sea level and evaporite deposition is limited, however, by the accuracy to which the timing of these events can be resolved. Recent advances in the development of astronomically tuned timescales in Mediterranean and deep-sea sequences have spurred a reassessment of the role of glacioeustatic changes in the evolution of the MSC (for a review, see Hilgen et al. [1999]).

Correlation of sedimentary cycles to astronomical changes in solar radiation has resulted in the development of very accurate timescales for the Neogene. Two approaches have been employed. Recovery of complete and continuous sections by the Ocean Drilling Program (ODP) have permitted orbital tuning of deep-sea sedimentary sequences through the Miocene [Shackleton et al., 1990, 1995a; Shackleton and Crowhurst, 1997; Tiedemann et al., 1994]. In the Mediterranean, correlation of sedimentary cycles with orbital precession has provided an astronomical timescale for the Plio-Pleistocene [Hilgen, 1991] and the late Miocene up to the base of the Messinian evaporites (for a review, see Hilgen et al. [1999]). A gap exists in the astronomical tuning of Mediterranean sequences in the late Messinian because marine sedimentation was interrupted by the isolation and desiccation of the Mediterranean during the MSC. This "Messinian gap" has been filled partially by astronomical tuning of sedimentary cycles in preevaporitic deposits in Spain, Italy, and Greece [Krijgsman et al., 1999a, 1999b; Hilgen and Krijgsman, 1999; Sierro et al., in press], and by correlation of evaporitic and terrestrial deposits in the Sorbas Basin of southern Spain [Krijgsman et al., 1999b; Sierro et al., 2001].

[1]Supporting data are available electronically at the World Data Center-A for Paleoclimatology, NOAA/NGDC, 325 Broadway, Boulder, CO 80303 (email: paleo@mail.ngdc.noaa.gov, URL: http://www.ngdc.noaa.gov/paleo/).

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In the deep sea, only two sedimentary sequences have been continuously recovered across the Miocene/Pliocene boundary and tuned astronomically. Shackleton et al. [1995a] developed a tuned timescale for the last 6 million years at ODP Site 846 in the eastern equatorial Pacific. At Site 926 on the Ceara Rise in the South Atlantic the sedimentary record was orbitally tuned from 14 to 5 Ma in the Miocene [Shackleton and Crowhurst, 1997]. During ODP Leg 162 to the North Atlantic, Site 982 (57°30.992'N, 15°52.001'W) was drilled in shallow water (1134 m) on the Rockall Plateau. Continuity of the composite sedimentary sequence was documented for the upper 255 meters below seafloor (mbsf), corresponding to ~7.5 Ma in the late Miocene, by correlation of core-logging signals [Shipboard Scientific Party, 1996]. Site 982 is therefore a good candidate for developing an orbitally tuned chronology for the late Miocene and early Pliocene.

We constructed an orbitally tuned timescale for Site 982 by correlating the record of gamma ray attenuation (GRA) bulk density to summer insolation at 65°N and the benthic $\delta^{18}O$ signal to orbital obliquity for the interval from 4.6 to 7.5 Ma. The astronomical timescale permits a direct correlation between proxy records in Site 982 and Messinian sedimentary cycles documented in the Mediterranean [Krijgsman et al., 1999a, 1999b; Hilgen and Krijgsman, 1999; Sierro et al., 2001]. The "tuned" time scale of Site 982 is used to reconstruct changes in late Miocene climate and global ice volume and assess the role they played in the isolation of the Mediterranean, deposition of the evaporites, and the reflooding of the basin at the Miocene/Pliocene boundary.

2. Methods

Cores within the Site 982 composite section were sampled at a constant interval of either 5 or 10 cm depending upon sedimentation rate, resulting in an average temporal sampling spacing of ~2500 years. Oxygen and carbon isotope ratios were measured on the benthic foraminifers *Planulina wuellerstorfi* and *Cibicidoides kullenbergi*. Previous studies have reported that the stable isotopic ratios of these two species are indistinguishable within analytical error. Measurements of paired analyses of both species from the same samples in Site 982 show a 1:1 correspondence for $\delta^{18}O$, but $\delta^{13}C$ values of *C. kullenbergi* diverge from *P. wuellerstorfi* at lower $\delta^{13}C$ values (Figure 1), suggesting that *C. kullenbergi* may sometimes live infaunally. For samples where paired analyses were made, oxygen isotope values were averaged for each sample, which significantly improved the signal to noise ratio in the upper Miocene sequence. For carbon isotopes, only $\delta^{13}C$ values measured on *P. wuellerstorfi* were used.

Samples between 122 and 138 meters composite depth (mcd) were analyzed for stable isotopes at Massachusetts Institute of Technology (MIT) and samples between 138 and 285 mcd were measured at the University of Florida (UF). Benthic foraminiferal tests were cleaned in an ultrasonic bath to remove fine-grained particles and soaked in either 15% H_2O_2 or roasted in a vacuum to remove organic matter. The foraminiferal calcite was reacted in a common acid bath of orthophosphoric acid at 90°C using a Micromass (formerly VG) Isocarb preparation system. Isotope ratios of purified CO_2 gas were measured online using Micromass Prism mass spectrometers at MIT and UF, respectively. Oxygen and carbon isotopes were also measured on bulk carbonate by grinding dried sediment to a homogenous powder and reacting untreated samples in orthophosphoric acid at 70°C using a Finnigan-MAT Kiel III carbonate preparation device. Evolved CO_2 gas was measured online with a Finnigan-MAT 252 mass spectrometer at UF. All isotope results are reported in standard delta notation

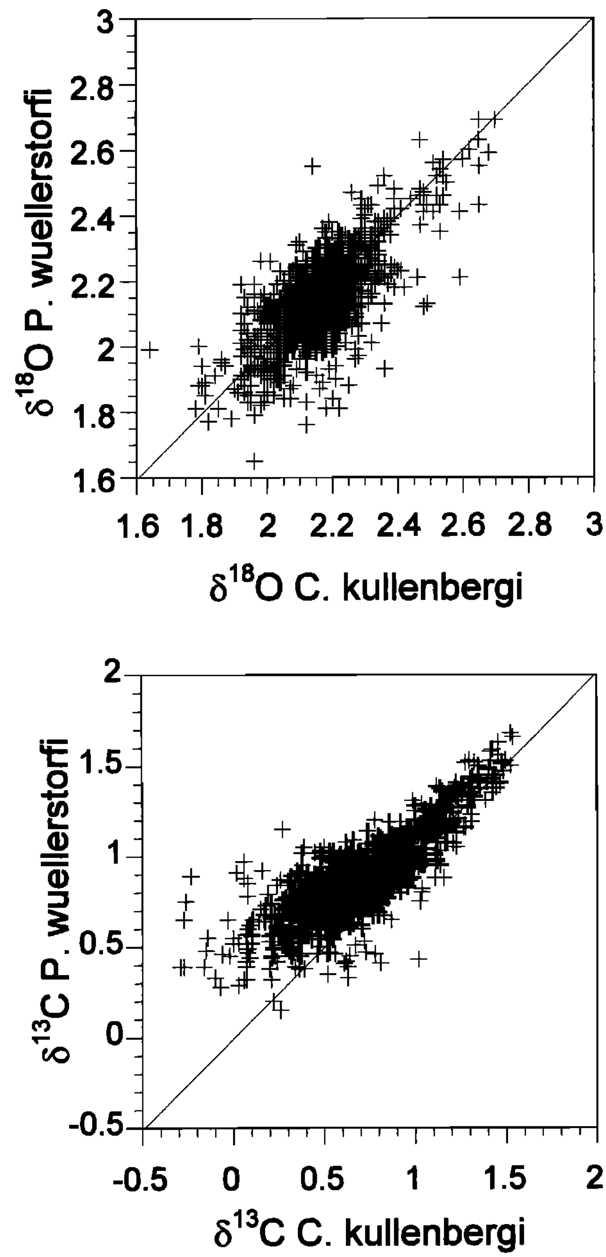

Figure 1. Oxygen and carbon isotope measurements of specimens of *P. wuellerstorfi and C. kullenbergi* from the same samples. Note that the trend follows a 1:1 relationship for $\delta^{18}O$ but diverges from 1:1 at lower $\delta^{13}C$ values, suggesting that *C. kullenbergi* may sometimes live infaunally.

Table 1. Depth-Age Points Used for Initial Chronology and Adjusted Ages of These Events After Tuning[a]

Datum	Depth (mcd)	Age (Ma)	Tuned age (Ma)
MIS NS6	128.09	4.713	4.778
MIS Si2	132.17	4.792	4.821
MIS Si4	136.55	4.871	4.854
MIS Si6	139.35	4.915	4.897
MIS T8	156.32	5.211	5.218
MIS TG12	172.78	5.540	5.506
MIS TG20	183.00	5.753	5.757
Event 4[b]	211.78	6.360	6.342
Event 3[c]	248.29	7.240	7.246
Event 1[d]	257.94	7.512	—

[a]Marine isotope stages (MIS) designations follow Shackleton et al. [1995b]. The ages for TG20 and TG12 were derived from Site 926 [Shackleton and Crowhurst, 1997], whereas all others were taken from Site 846 [Shackleton et al., 1995b] and adjusted by adding 41 kyr corresponding to one obliquity cycle. Foraminiferal events are described by Sierro et al. [1993], and the astronomical ages were derived from Hilgen et al. [1995].

[b]First common occurrence of dextral forms of *Neogloboquadrina acostaensis*

[c]Replacement of the *Globorotalia menardii* group by the *Globorotalia miotumida* group

[d]Last occurrence of *G. menardii* sinistral

relative to Vienna Pee Dee Belemnite (VPDB) [Coplen, 1996].[1] Analytical precision for all isotope analyses was better than ±0.1‰. Weight percent carbonate was measured by coulometric titration using a UIC Inc. Model 5240 Total Inorganic Carbon (TIC) preparation system and Model 5011 coulometer [Englemann et al., 1985]. GRA bulk density measurements were made aboard Leg 177 using a GEOTEK multi-sensor core logger [Shipboard Scientific Party, 1996].

3. Chronology and Astronomical Tuning

No polarity reversal stratigraphy is available for the upper Miocene to lower Pliocene section of Site 982 because magnetization intensities were low and within the noise level of the shipboard pass-through cryogenic magnetometer [Shipboard Scientific Party, 1996]. An initial chronology for Site 982 was constructed using prominent marine isotope stages (Table 1 and Figure 2) [Shackleton et al., 1995] and three planktic foraminiferal events that have been calibrated previously to the astronomical time scale (Table 1) [Sierro et al., 2001; Hilgen et al., 1995]. For Pliocene isotope stages we used the published ages of Shackleton et al. [1995b] adjusted by 41 kyr, corresponding to one obliquity cycle [Lourens et al., 1996]. Because Shackleton et al. [1995a] used the astronomical solution of Berger and Loutre [1991] and obtained a much younger estimate for the base of the Gilbert than did Krijgsman et al. [1999a], we used the ages for late Miocene isotope stages TG20 and TG12 from Site 926 [Shackleton and Crowhurst, 1997]. The tuning of Site 926 signals used the orbital solution of La90 (1,1), and the astronomically tuned chronology is consistent with Hilgen et al. [1995] in the late Miocene [Shackleton and Crowhurst, 1997].

Recognition in Site 982 of the coiling change in *Neogloboquadrina acostaensis* (from sinistral to dextral; Figure 2) is an important datum because it has been found in other North Atlantic sites in the middle of Chron C3An.1r and is synchronous with the same event in the Mediterranean [Hooper and Weaver, 1987; Hodell et al., 1994; Sierro et al., 2001]. The coiling change has been dated at 6.36 Ma throughout the Mediterranean and occurs 17 precessional cycles below onset of evaporite deposition [Hilgen and Krijgsman, 1999]. The replacement of the *Globorotalia menardii* group by the *Globorotalia miotumida* group (event 3 of Sierro et al. [1993]) occurs near the Tortonian/Messinian boundary and is dated at 7.24 Ma throughout the Mediterranean. Last, the LaO of *G.*

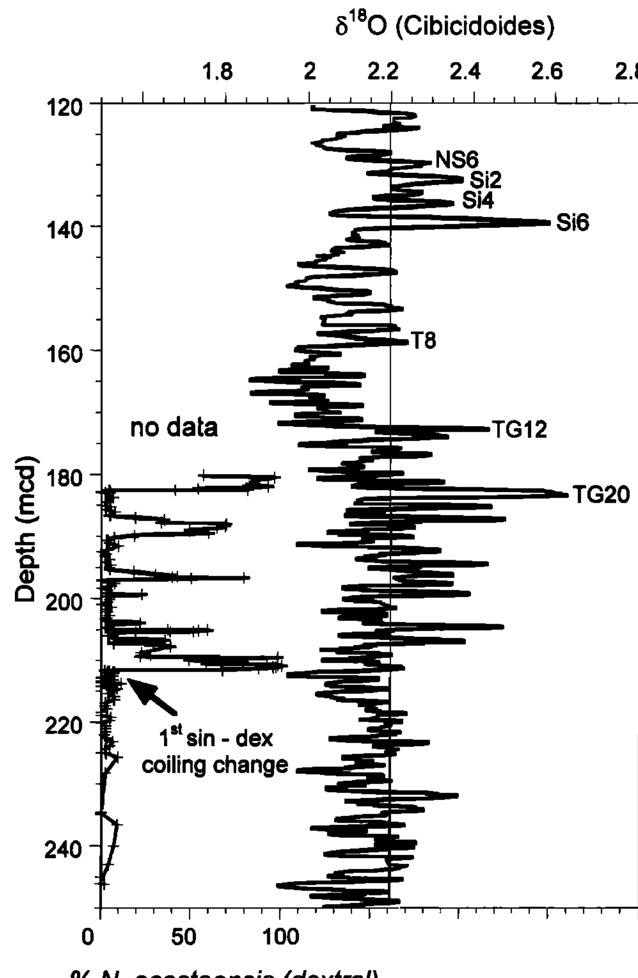

Figure 2. (left) Coiling ratio of *Neogloboquadrina acostaensis* expressed as percent dextral to total specimens. The first coiling change from dominantly sinistral to dextral forms occurs at 211.78 meters composite depth (mcd) and has been dated in the Mediterranean at 6.36 Ma [Hilgen and Krijgsman, 1999]. (right) Three-point running mean of the benthic oxygen isotopic signal showing prominent isotope stages identified by Shackleton et al. [1995b] and used in constructing the preliminary age model (see Table 1).

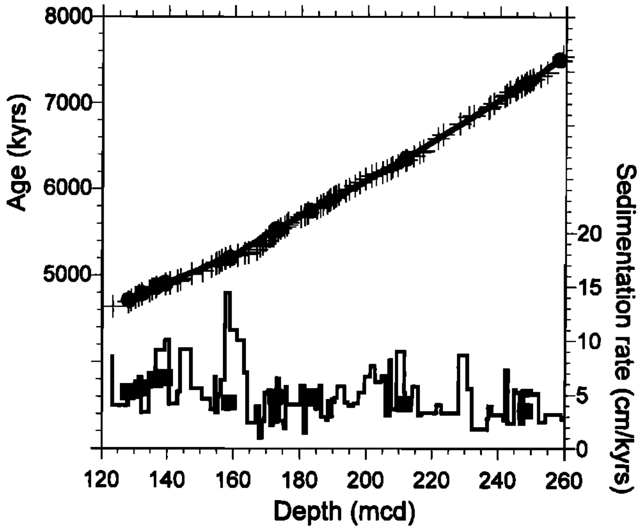

Figure 3. Age-depth plot of biostratigraphic and isotopic control points (solid circles) used to construct preliminary age model of Site 982 and age-depth pointers (crosses) derived after tuning the record. Interval sedimentation rates are shown for the initial calibration points (solid squares) and tuned record (solid line).

menardii (sinistral) occurs just below the base of the composite section at Site 982 and is dated at 7.51 Ma.

As a starting point for astronomical tuning, depths (mcd) in Site 982 were converted to age by linearly interpolating between age-depth pairs in Table 1. Sedimentation rates average 4.7 cm/kyr in the late Miocene and early Pliocene, and interval sedimentation rates of the tuned record vary by an order of magnitude from ~1 to 10 cm/kyr (Figure 3). The benthic $\delta^{18}O$ and GRA bulk density signals were tuned to the La90 $_{(1,1)}$ astronomical solution [Laskar et al., 1993] by filtering the benthic $\delta^{18}O$ signal at 41 kyr and matching it to obliquity and filtering the GRA signal at 21 kyr and matching it to summer insolation at 65°N (Figures 4a-4c). The tuning was done iteratively in order to obtain the best fit for both $\delta^{18}O$ and GRA signals. Previous studies have shown the benthic $\delta^{18}O$ signal to be dominated by the 41-kyr obliquity cycle in the late Miocene and Pliocene [Hodell et al., 1994; Shackleton et al., 1995b; Shackleton and Hall, 1997; Tiedemann et al., 1994; Tiedemann and Franz, 1987].

The GRA bulk density record in Site 982 is correlated with wt% carbonate content (Figure 5). In general, higher bulk density and carbonate content coincide with lower values of bulk carbonate and benthic $\delta^{18}O$, indicating

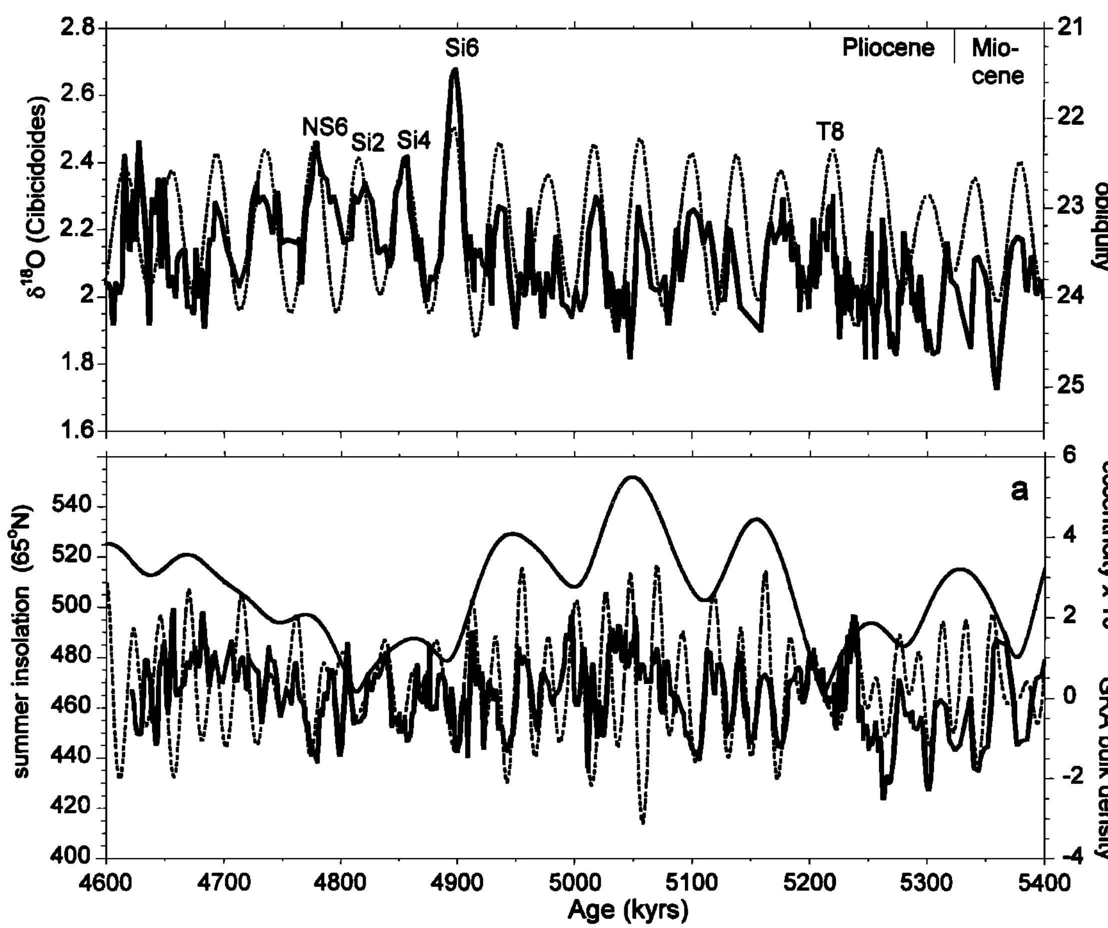

Figure 4. Benthic $\delta^{18}O$ signal tuned to obliquity and gamma ray attenuation (GRA) bulk density to summer insolation at 65°N using the La90$_{(1,1)}$ astronomical solution of Laskar et al. [1993]. The GRA bulk density record is normalized to unit variance, and eccentricity values have been multiplied by 100 for illustration purposes. Prominent oxygen isotope stages are identified in Figures 4a and 4b according to Shackleton et al. [1995b]. UA16 to UA34, Y1 to Y14, and SI to SIII refer to cycle numbers in the upper Abad, Yesares, and Sorbas Members of the Sorbas Basin of southern Spain [Sierro et al., 2001; Krijgsman et al., 2001].

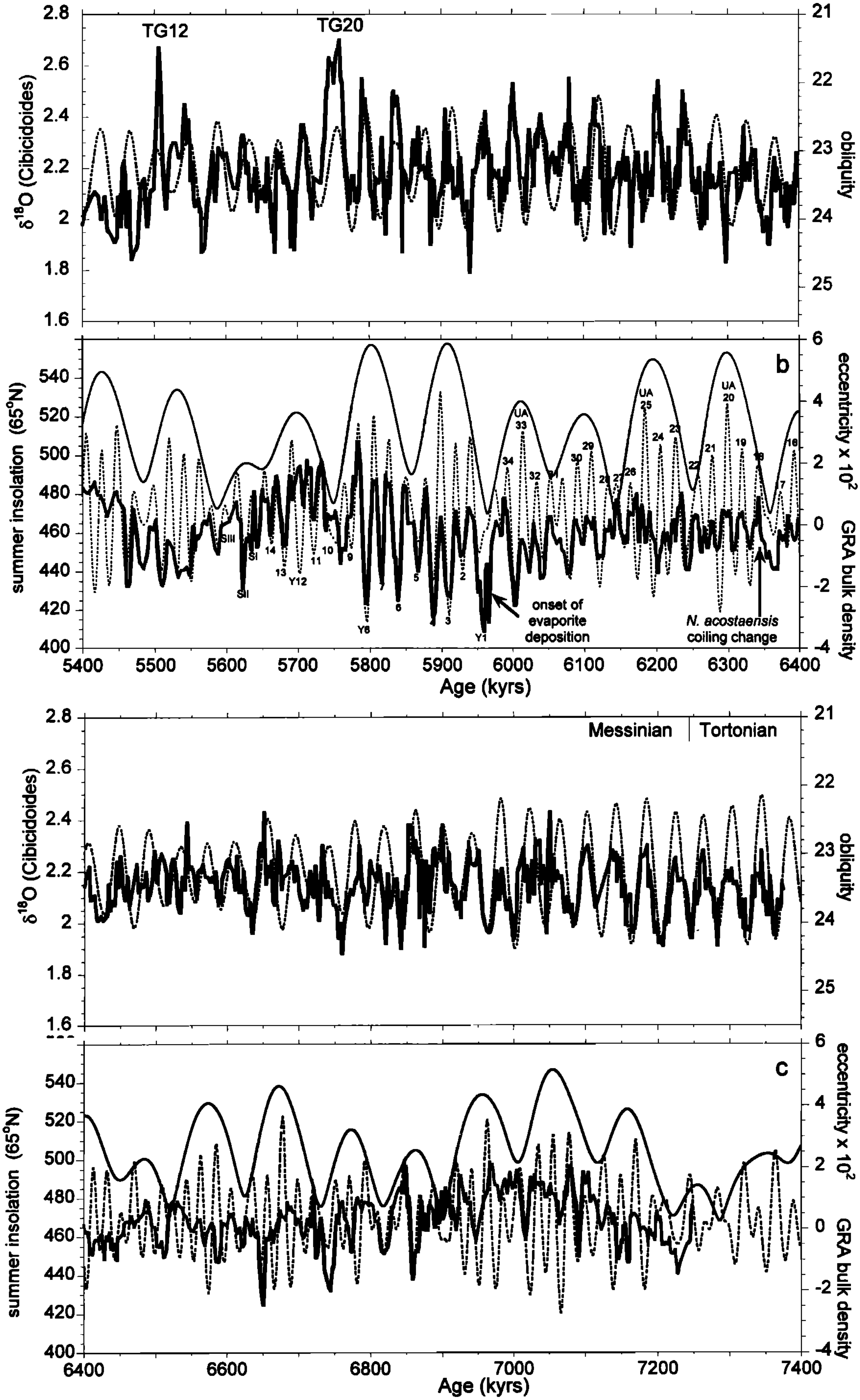

Figure 4. (continued)

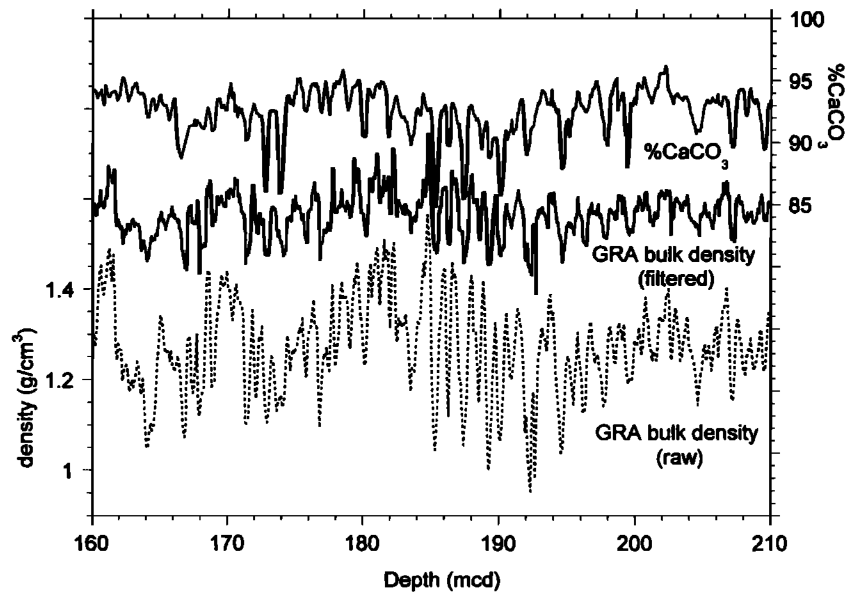

Figure 5. Comparison of GRA bulk density signal measured on whole-cores using a GEOTEK multisensor logger before (dashed line) and after (solid line) filtering to remove a 1.5-m cycle due to section breaks. The units of the filtered GRA bulk density signal are arbitrary and have been scaled to match the amplitude of the weight percent carbonate record. Note the general correspondence between carbonate content and GRA bulk density.

warmer surface and deep waters. We use this relationship as justification for tuning the GRA bulk density record to summer insolation at 65°N. Baumann and Huber [1999] adopted a similar approach for the Plio-Pleistocene of Site 982 by tuning the unfiltered carbonate record to high-latitude summer insolation. In addition, Beaufort and Aubry [1990] found major fluctuations in the nannofossil assemblages in the late Miocene of nearby Deep Sea Drilling Project (DSDP) Site 552 on the Rockall Bank (56°N, 10°S) that were dominated by the 20- and 100-kyr cycles of orbital precession and eccentricity, respectively. Sprovieri et al. [1999] tuned the planktic oxygen isotope and nannofossil abundance variations in Site 552 to insolation using $La90_{(1,1)}$, but continuity of the section cannot be assured because it was recovered in a single hole.

The power spectrum of the GRA bulk density signal in the depth domain revealed a peak at 1.5 m that is not seen in the power spectrum of percent $CaCO_3$. This cycle may be an artifact of the fact that each 9.5-m ODP core is cut into six 1.5-m sections. Partially filled core liners can produce low GRA bulk density values near the tops and bottoms of sections when cores are analyzed with the GEOTEK multi sensor logger. Although this artifact was corrected for on board ship by eliminating data from the first and last 5 cm of each section [Shipboard Scientific Party, 1996], this treatment apparently did not eliminate the effect entirely. To remove the 1.5-m cycle from the record, the GRA bulk density record was filtered in the depth domain at 1.5 m, and the filtered signal was subtracted before converting the signal to an age scale. The corrected GRA bulk density signal was then correlated to summer insolation at 65°N. Filtering the 1.5-m cycle from the GRA bulk density signal does not alter the relationship between the GRA bulk density and $CaCO_3$ signals (Figure 5). In fact, the correlation of the GRA bulk density and insolation records would be no different if we had used the filtered versus unfiltered GRA bulk density record. We used the GRA bulk density signal for tuning instead of the percent $CaCO_3$ record because core logging was done at continuous 1-cm increments as opposed to 5-cm spacing of percent $CaCO_3$ measurements. In addition, the two records are not identical in that the GRA bulk density signal has more variation at precessional frequencies than does percent $CaCO_3$ (Figure 6).

The accuracy of the tuning varies in different parts of the record as the amplitude and frequency of the oxygen isotope and GRA density signals change. For example, the tuning between ~5.2 Ma (isotope stage T8) and 6.4 Ma (*N. acostaensis* coiling change) is exceptionally good because the GRA signal contains a strong precession and obliquity component (Figures 4b and 6a). In addition, the benthic $\delta^{18}O$ signal is also strong in this interval and dominated by obliquity (Figure 6b). This critical period includes the MSC that is the focus of this paper.

Continuity of the composite section was confirmed for the upper 255 mcd of Site 982 only [Shipboard Scientific Party, 1996], and below this level the section was recovered in a single hole (982B). The chronology below 255 mcd was derived by correlating the benthic carbon isotope records of Hole 982B and the Salé Briqueterie Section, northwestern Morocco [Hodell et al., 1994]. For this purpose, the paleomagnetic reversal stratigraphy of the Salé Section was recalculated using the ages given by Krijgsman et al. [1999a] to the base of Chron C3An.2n and Hilgen et al. [1995] thereafter. The carbon isotopic records of Site 982 and Salé are easily correlated between ~9 and 7 because of a maximum in $\delta^{13}C$ in Chron C4n followed by the late Miocene carbon shift [Hodell et al., 1994].

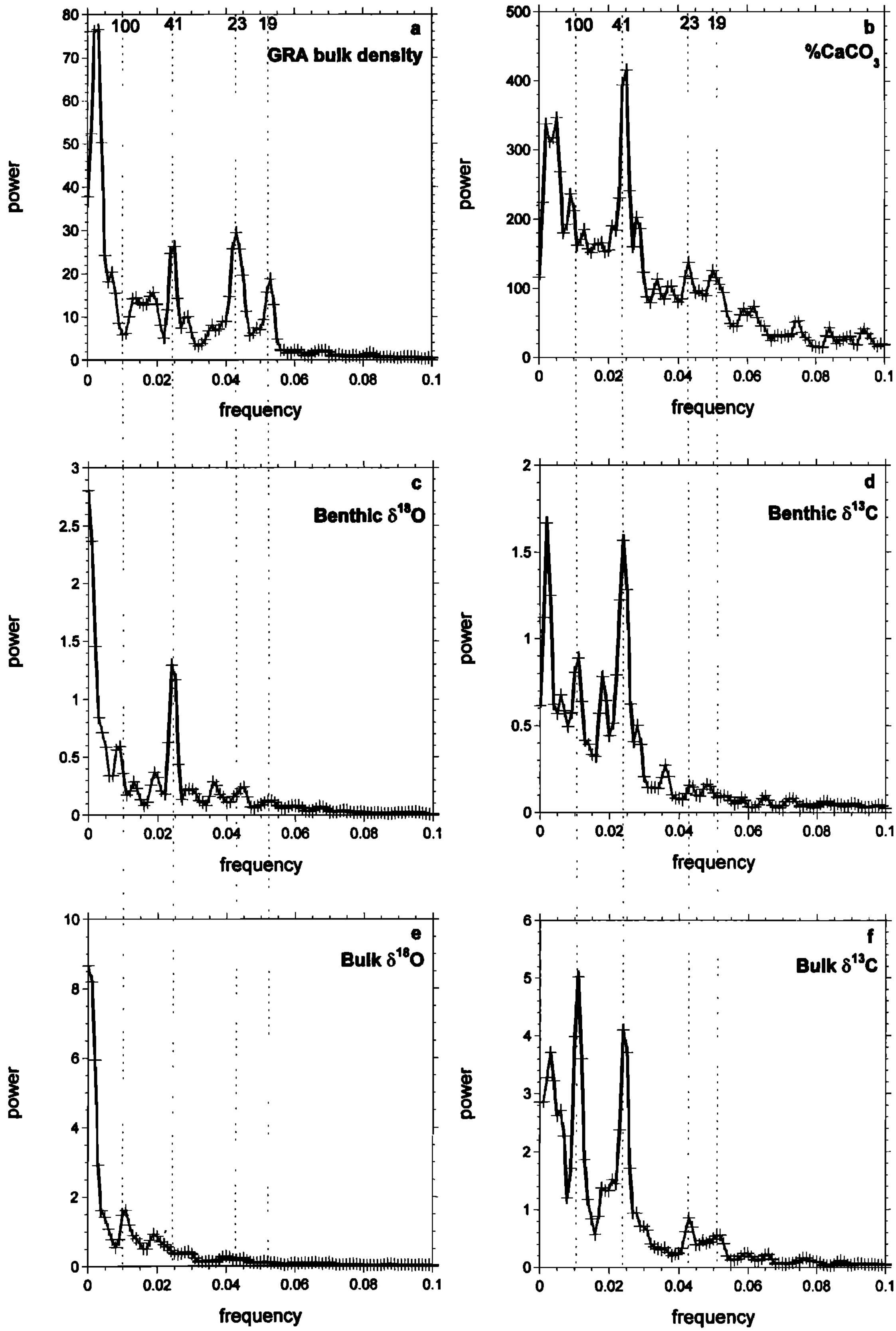

Figure 6. Spectral analysis of proxy signals in Site 982 for the interval between 5.21 and 6.36 Ma, i.e., between the coiling direction change in *N. acostaensis* and marine isotope stage (MIS) T8. All signals were linearly detrended, normalized to unit variance, and re-sampled at a constant time interval of 1 kyr. Spectra were calculated by Analyseries [Paillard et al., 1996] using the Blackman-Tukey method with one-third lag and Barlett window

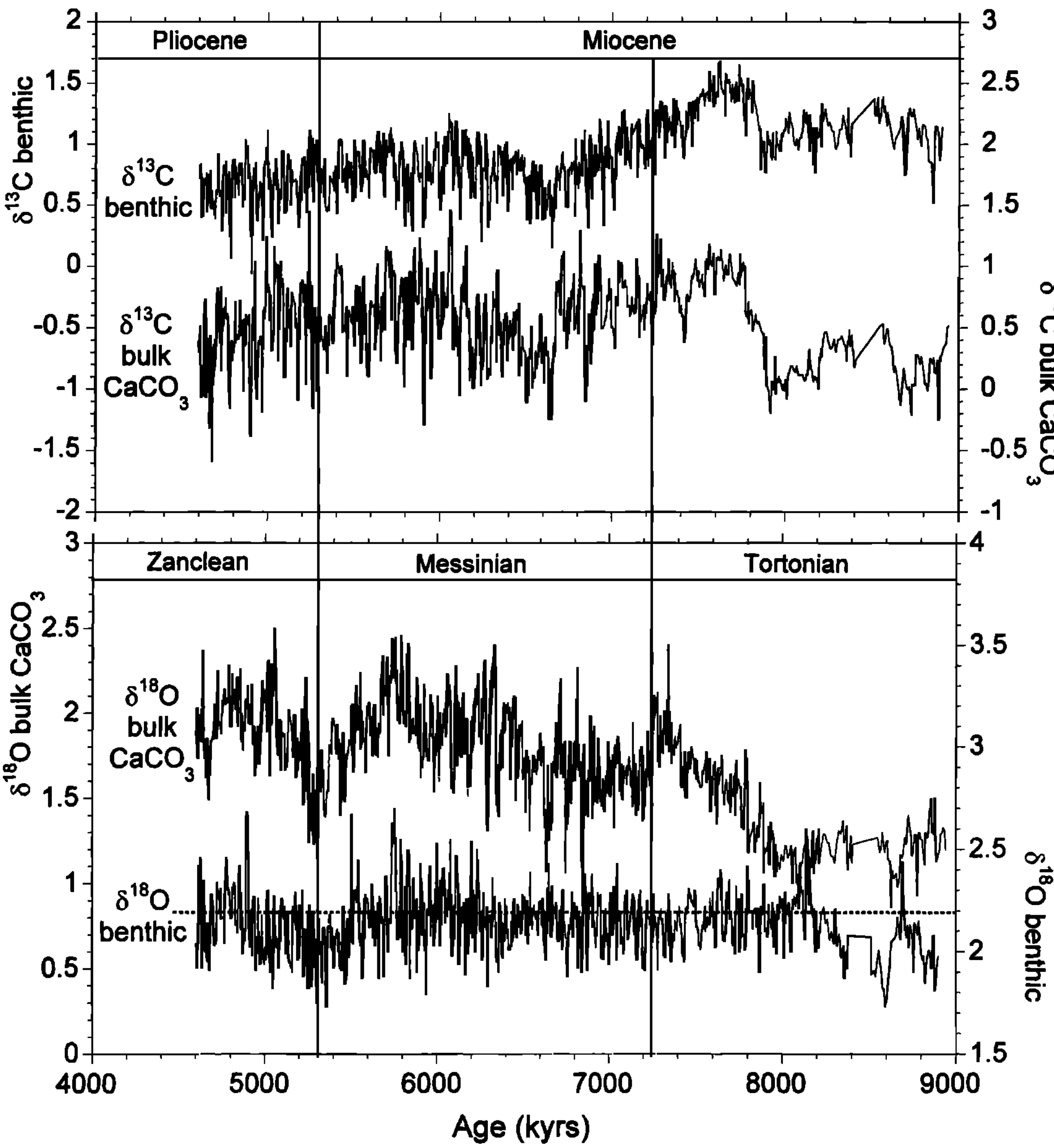

Figure 7. (top) Carbon isotopic record of benthic foraminifera and bulk carbonate in Site 982. (bottom) Oxygen isotopic record of benthic foraminifera and bulk carbonate in Site 982. The dashed horizontal line through the benthic $\delta^{18}O$ signal is the measured Holocene value at the top of Site 982 [Venz et al., 1999].

4. Results

Benthic $\delta^{18}O$ results are compared to Holocene $\delta^{18}O$ values (2.2‰) measured at the top of Site 982 [Venz et al., 1999]. The mean of the $\delta^{18}O$ signal over the period from 4.6 to 9 Ma was fairly close to the Holocene value (Figure 7). There are two distinct intervals between 6.2 and 5.5 Ma and from 4.9 to 4.6 Ma when peak $\delta^{18}O$ values were significantly greater than the Holocene by up to 0.5‰. In contrast, the interval from 5.5 to 5.2 Ma was marked by lower $\delta^{18}O$ values than the Holocene by up to 0.3‰. The $\delta^{18}O$ record of bulk carbonate shows substantially more variability than the benthic $\delta^{18}O$ signal (Figure 7). An increase in mean $\delta^{18}O$ values occurs between 7.9 and 7.3 Ma followed by a drop at the Tortonian/Messinian boundary. A trend toward higher bulk $\delta^{18}O$ values begins at 6.3 Ma, and the greatest values are found between ~5.8 and 5.7 Ma. This peak is followed by a progressive decrease reaching minimum values near the Miocene/Pliocene boundary between ~5.2 and 5.3 Ma. Bulk $\delta^{18}O$ values increase again in two steps at 5.2 to 5.06 Ma in the early Pliocene.

The $\delta^{13}C$ records of both benthic foraminifers and bulk carbonate show an increase in the late Tortonian at ~7.8 Ma and reach maximum values between 7.8 and 7.6 Ma (Figure 7). Between ~7.6 and 6.8 Ma, benthic $\delta^{13}C$ values decrease across the Tortonian/Messinian boundary, marking the well-known late Miocene carbon shift. The carbon shift is not obvious in the bulk $\delta^{13}C$ record except for an abrupt decrease at 6.67 Ma near the end of the benthic $\delta^{13}C$ decrease. From 6.4 to 4.6 Ma both records are marked by high variability with no pronounced change in mean $\delta^{13}C$ values.

5. Discussion

5.1. Comparison of Site 982 and Salé Briqueterie Section (Morocco)

Hodell et al. [1994] presented a detailed benthic stable isotopic record from the Salé Briqueterie drill core in northwest Morocco located at the western end of the Rifian Corridor, which connected the Mediterranean and Atlantic during the late Miocene. Comparison of the Site 982 and Salé $\delta^{18}O$ records permits us to identify those isotope changes that are localized to the Mediterranean gateway and those that are characteristic of the North Atlantic. The Morocco record is marked by a distinct increase in benthic $\delta^{18}O$ values across the Tortonian/Messinian boundary from ~7.3 to 6.9 Ma (Figure 8a). This $\delta^{18}O$ increase has been interpreted to reflect a reversal in deepwater flow through

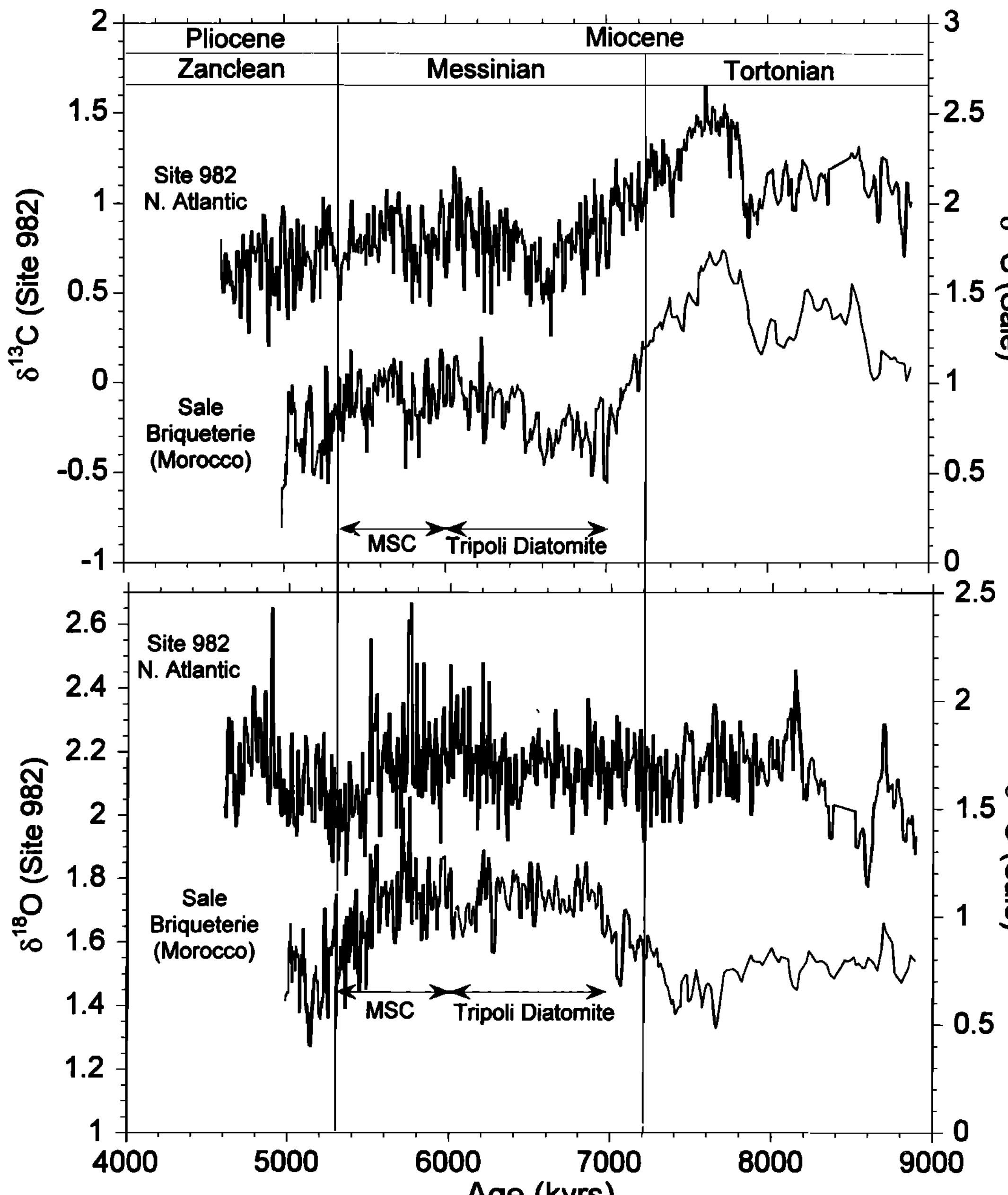

Figure 8. (top) Comparison of the carbon isotopic records from Site 982 and the Salé Briqueterie section of northwestern Morocco. Note the late Miocene carbon shift that occurred across the Tortonian/Messinian boundary in both records. (bottom) Comparison of the oxygen isotopic records from Site 982 and the Salé Briqueterie section of northwestern Morocco [Hodell et al., 1994]. Note the increase in $\delta^{18}O$ values across the Tortonian/Messinian boundary in Morocco that is not mirrored in the North Atlantic record. The lines with double arrows show the age ranges of the Tripoli diatomite in Sicily and the Messinian salinity crisis (MSC) according to the chronology of Hilgen and Krijgsman [1999] and Krijgsman et al. [1999a], respectively.

the Rifian Corridor, whereby warm waters of Mediterranean origin were replaced by cold intermediate waters of Atlantic origin [Hodell et al., 1989; Benson et al., 1991; Hodell et al., 1994]. Hodell et al. [1994] also suggested that part of the Salé $\delta^{18}O$ increase may reflect an increase in continental ice volume that lowered sea level, but this interpretation is not supported by the Site 982 $\delta^{18}O$ data because no increase occurred in the North Atlantic near the Tortonian/Messinian boundary that would indicate a change in intermediate water temperatures or continental ice volume. Contrary to previous speculation, the early restriction of the Betic and Rifian Straits was not associated with a major glacioeustatic lowering of sea level and therefore was most probably tectonic in origin. Tectonic uplift in both the Betic and Rifian Corridors has been documented in the latest Tortonian and earliest Messinian [Weijermars, 1988; Martin and Braga, 1996; Krijgsman et al., 1999b, 2001].

The $\delta^{18}O$ increase in Morocco between ~7.3 and 6.9 Ma represents cooling or increased salinity that resulted from a change in circulation in the Rifian gateway and the establishment of a negative water balance in the Mediterranean. In the Mediterranean, the period from 7.16 to 6.8 Ma is marked by a major shift in benthic foraminiferal assemblages, whereby species indicative of normal marine conditions are replaced by taxa indicative of lower-oxygen and higher-salinity conditions [Kouwenhoven et al., 1999]. This change is accompanied by an increase in Mediterranean planktic and benthic $\delta^{18}O$, including a pronounced increase just after 6.8 Ma [Glacon et al., 1990; Kouwenhoven et al., 1999]. These changes document the progressive tectonic isolation of the Mediterranean and are not associated with any significant changes in the amplitude of the benthic $\delta^{18}O$ signal at Site 982 (Figure 7).

The carbon isotopic records of Site 982 and Salé are nearly identical (Figure 8), suggesting that the benthic $\delta^{13}C$ signal in Morocco is an intermediate-water Atlantic signal that was not significantly influenced by processes in the Mediterranean. The carbon shift is also recorded in Mediterranean planktic and benthic foraminifers between ~7.16 and 6.85 Ma, but the magnitude is much greater in the Mediterranean than in the Atlantic [Kouwenhoven et al., 1999]. As noted previously [Keigwin, 1987; Muller et al., 1991], the late Miocene carbon shift in Site 982 was not associated with an increase in benthic $\delta^{18}O$ in the open ocean, thereby precluding sea level as a forcing mechanism. The carbon shift is synchronous, however, with the change in Mediterranean circulation inferred from the benthic $\delta^{18}O$ increase in Morocco. Because it is unlikely that a change in the production of Mediterranean Overflow Water (MOW) could lead to a global change in the $\delta^{13}C$ of the TCO_2 of both surface and deep waters, the reversal in circulation in the Mediterranean and late Miocene carbon shift were probably not causally related. The shift in the Mediterranean water balance near the Tortonian/Messinian boundary can be explained by a sudden increase in evaporation or decrease in continental drainage. Palynological evidence indicates a trend toward drier Mediterranean climate in the late Miocene [Suc and Bessais, 1990], which may have triggered the change in the Mediterranean's water budget and circulation. The same climatic drying may have also decreased terrestrial biomass and favored the expansion of C_4 habitats, resulting in a net transfer of carbon from terrestrial to oceanic pools and the late Miocene carbon shift [Hodell et al., 1994]. Alternatively, it is possible that both the carbon shift and constriction of Mediterranean gateways were related to a global tectonic event. Krijgsman et al. [1999] suggested that tectonic closure of the Mediterranean may have been related to global changes in spreading rate and plate motion, such as those occurring in the Pacific in the late Miocene [Cox and Engebretson, 1985]. Muller et al. [1991] proposed that the carbon shift was associated with tectonic uplift in the circum-Pacific that eliminated "Monterey-type basins" as sinks for organic carbon in the late Miocene [Teng and Gorsline, 1989; Barron, 1986].

5.2. Latest Miocene Glaciation

Reports of a major cooling episode and increase in global ice volume in the late Miocene are widespread (for a review, see Hodell and Kennett [1986]). The $\delta^{18}O$ record of Site 982 demonstrates that the period from 6.26 (near base of Chron C3An.1n) to 5.5 Ma in the latest Miocene was marked by 18 glacial-to-interglacial oscillations that were dominantly controlled by the 41-kyr cycle of obliquity (Figure 9). The most prominent glacial stages during this interval were TG20 at 5.75 Ma and TG12 at 5.51 when $\delta^{18}O$ values exceeded the Holocene by ~0.5‰. Most glacial stages in the late Miocene were only 0.3‰ greater than Holocene values, however. Reports of ice-rafted debris in sediments off Greenland and in the Norwegian Sea suggest that glaciers were large enough to reach sea level during the late Miocene [Jansen and Sjoholm, 1991; Larsen et al., 1994].

The magnitude of sea level lowering associated with late Miocene glaciations is difficult to determine from $\delta^{18}O$ data alone because of the combined effects of temperature and ice volume in the signal. From a study of the Niue Atoll, Aharon et al. [1993] estimated sea level changes of ~10 m amplitude beginning at 6.14 Ma reaching an amplitude of at least 30 m at 5.26 Ma (timescale of Hilgen [1991]). Because the dating was done by strontium isotope stratigraphy, which has considerable error (±0.5 Ma) in the late Miocene [McKenzie et al., 1988], the absolute ages from the Niue Atoll cannot be compared with the Site 982 record, but the interval has been shown to correlate to the MSC [Aharon et al., 1993]. Braga and Martin [1996] estimated sea level changes of the order of tens of meters in the Sorbas Basin by studying variations in the elevation of reef facies. These estimated sea level variations of 10-30 m are consistent with the amplitude of the benthic $\delta^{18}O$ changes in Site 982 (Figure 9). We cannot exclude the possibility, however, that the observed benthic $\delta^{18}O$ variability reflects changes in intermediate water temperature at Site 982 and are therefore unrelated to global ice volume.

5.3. Correlation of Site 982 to Mediterranean Sections

The tuning of the Site 982 record to summer insolation permits a direct bed-to-bed correlation to the preevaporitic

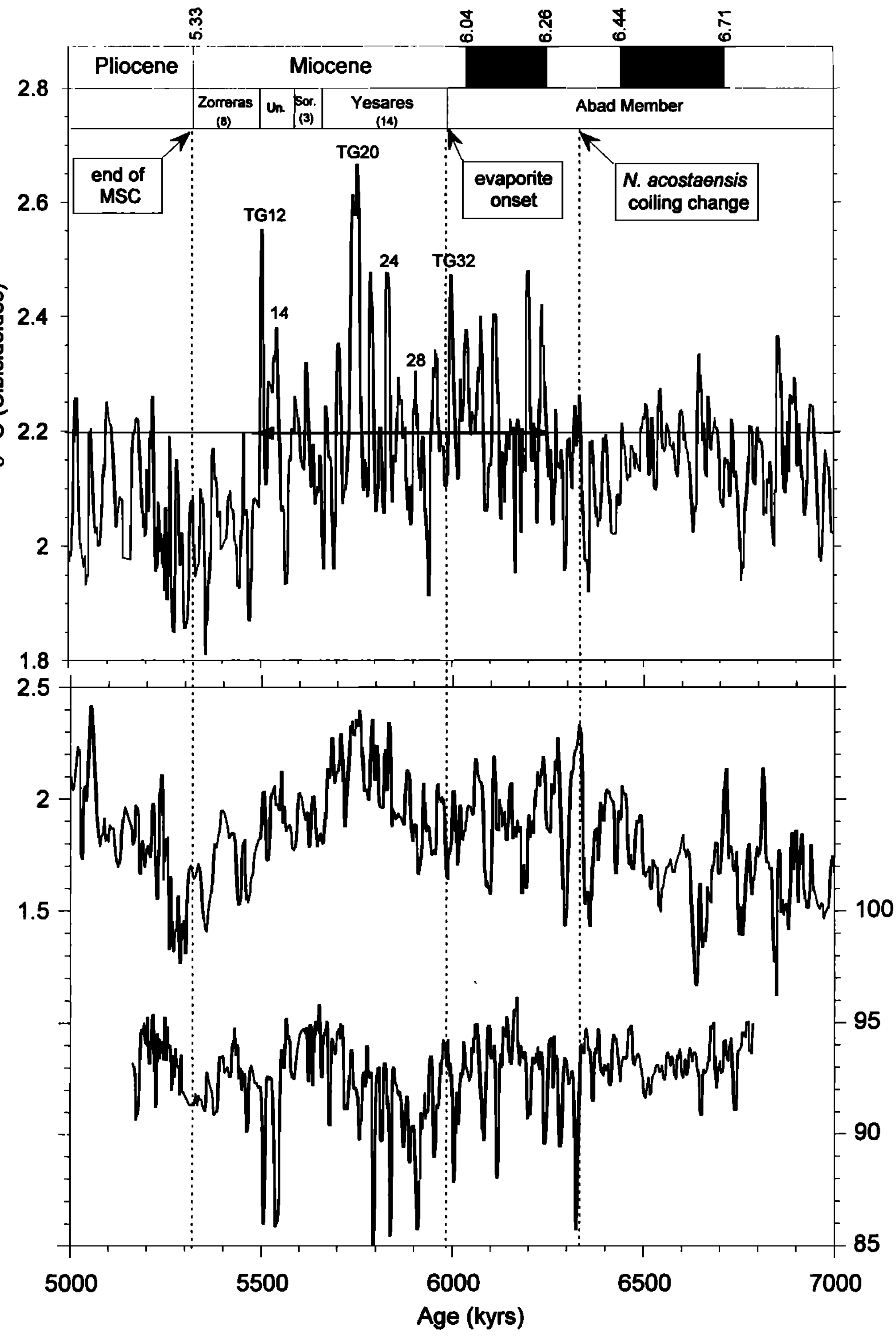

Figure 9. Three-point running average of the oxygen isotope signals measured on (top) benthic foraminifera and (bottom) bulk carbonate and weight percent $CaCO_3$ in Site 982. Horizontal line on top panel represents the measured Holocene $\delta^{18}O$ value at the top of Site 982. Thick horizontal line with arrows indicates the age range of the latest Miocene glaciation. Interpretation of the stratigraphy and chronology of the Sorbas Basin according to Krijgsman et al. [2001] is shown at the top along with inferred geomagnetic polarity stratigraphy.

Messinian sections in the Mediterranean (Figure 4b). Studies of these sections in Spain, Italy, and Greece have resulted in the development of a cyclostratigraphic framework for the Mediterranean that is tied to high-resolution planktic foraminiferal biostratigraphy [Krijgsman et al., 1999a; Sierro et al., 2001]. We compare the isotopic records from Site 982 with the cyclostratigraphy of the Sorbas Basin in the western Mediterranean because this section is only 350 km from the North Atlantic and therefore is most likely to be affected by open-ocean processes prior to the isolation of the basin.

The uppermost biostratigraphic event that can be identified in both Site 982 and Sorbas is the first change in the coiling direction of *N. acostaensis* (from sinistral to dextral; Figures 2 and 4b) dated at 6.36 Ma [Krijgsman et al., 1999]. This event has been recognized in other North Atlantic sites as occurring in the middle of Chron C3An.1r and is synchronous with the event in the Mediterranean [Hooper and Weaver, 1987; Hodell et al., 1994]. In Sicily, the coiling change is recognized in Tripoli diatomite cycle T32 or 17 cycles below the top of the section, which marks the onset of evaporite deposition [Hilgen and Krijgsman, 1999]. In the Sorbas Basin the coiling direction change is recognized in cycle UA18. Planktic foraminifera disappear in the Upper Abad Section beginning with cycle UA17 just above the coiling change in *N. acostaensis*. In Site 982 the coiling change is associated with a distinct decrease in weight percent $CaCO_3$ and increase in the $\delta^{18}O$ of bulk carbonate (Figure 9), which may indicate cooling of surface waters or, alternatively, a change in nannofossil assemblages [Dudley and Goodney, 1979; Paull and Thierstein, 1987].

About 100 kyr above the coiling change in *N. acostaensis* near the base of Chron C3An.1n at 6.26 Ma, there is a distinct increase in the amplitude of the benthic $\delta^{18}O$ signal at Site 982 (Figure 9). This time marks the start of the latest Miocene glaciation that consists of 18 glacial-to-interglacial cycles that were dominantly controlled by high-latitude insolation changes related to Earth's obliquity (Figure 4). The increase in $\delta^{18}O$ at 6.26 Ma correlates with the transition from cycle T36 to T37 in Sicily and UA22 to UA23 in Sorbas Basin (Figure 4b). This cycle also coincides with a major slumping event seen in all sections of the Upper Abad Member in the Sorbas Basin [Sierro et al., 2001], and may correlate to an unconformity that separates the bioherm unit from the fringing reef unit in marginal deposits of the Sorbas Basin [Martin and Braga, 1994, 1996]. The coiling change in *N. acostaensis* has been found in the bioherm unit and is consistent with the unconformity at the base of the fringing reef unit correlating with slumping in the deeper parts of the Sorbas Basin [Riding et al., 1998]. This erosional surface and slumping may have been related to sea level lowstands during the interval of increased $\delta^{18}O$ amplitude beginning at 6.26 Ma. There is also a change in the nature of the cyclic sedimentation at this level in the Abad Member of Sorbas, which is marked by the common occurrence of indurated layers or paper shales in the cyclic bedding [Sierro et al., 2001].

Because the cycles in the Sorbas Basin are dominated by precession (as they also are in Sicily and Greece), whereas the benthic $\delta^{18}O$ is dominated by obliquity (Figure 9), Sierro et al. [1999] concluded that glacioeustatic changes were not primarily responsible for the faunal and sedimentary cycles in Sorbas. Similar conclusions have been reached for preevaporitic Messinian sections in Sicily and Greece [Krijgsman et al., 1999a; Hilgen and Krijgsman, 1999]. The top of the Abad Member in the Sorbas Basin (cycle UA34) is dated at 5.96 ±0.02 Ma and is time equivalent to the onset of evaporites or evaporitic limestones in northern Italy, Sicily, and Greece [Krijgsman et al., 1999a], marking the initiation of the MSC. There is no significant change in the $\delta^{18}O$ records of either benthic foraminifers or bulk carbonate at 5.96 Ma in Site 982 (Figure 9), suggesting that glacioeustatic lowering of sea level was not the cause of the initiation of deposition of the lower evaporite unit. However, there are prominent lows in GRA bulk density at 5.96 Ma and several minima in weight percent $CaCO_3$ associated with the base of the lower evaporite unit (Figures 4b and 9). Increased dissolution is a common feature of late Miocene records in the Atlantic and may reflect a shoaling of the lysocline in response to rapid extraction of calcium carbonates and sulfates during the MSC [Thunell et al., 1987; Muller et al., 1991].

Until recently, a gap has existed in the astronomical tuning of Mediterranean sequences in the late Messinian because marine sedimentation was interrupted by the MSC. This "Messinian gap" has been filled partially by astronomical tuning of evaporitic and terrestrial deposits in the Sorbas Basin of southern Spain [Krijgsman et al., 1999b; Sierro et al., 2001]. Correlation of these sequences rests on the interpretation and enumeration of cycles within the Yesares, Sorbas, and Zorreras Members of the Sorbas Basin [Krijgsman et al., 1999, 2001]. Krijgsman et al. [1999] suggested that the Yesares and Sorbas Members represent the lower evaporites and consist of 16-17 cycles that were deposited between 5.96 and 5.59 Ma (Figure 9). An unconformity at the top of the Sorbas Member between 5.59 and 5.5 Ma marks the boundary between the lower and upper evaporites when the Mediterranean was desiccated and erosion incised canyons along the Mediterranean margins. The Zorreras Member corresponds to the upper evaporites and consists of eight cycles deposited between 5.5 and 5.33 Ma.

One of the most prominent benthic $\delta^{18}O$ increases of the latest Miocene was stage TG20 at 5.75 Ma when values rose to ~0.5‰ greater than the Holocene (Figure 9). Shackleton et al. [1995b] suggested that the sea level fall associated with this event may have triggered the complete isolation of the Mediterranean from the world ocean. Krijgsman et al. [1999a] argued this was not the case because according to their chronology, TG20 (5.75 Ma) postdates the onset of evaporite deposition (5.96 Ma) and predates the total isolation event (~5.5 Ma) by ~200 kyr (Figure 9). According to cyclostatigraphic correlation of Sorbas Basin and Site 982, stage TG20 coincides with evaporite cycles 10, 11, and 12 in the Yesares Member, which consist of thin evaporitic limestones instead of gypsum [Krijgsman et al., 2001].

Following the deposition of the lower evaporites, the Mediterranean became completely isolated from the Atlantic, resulting in erosion and channelization of its continental shelves and slopes. Krijgsman et al. [1999a] suggested the intra-Messinian hiatus lasted for ~100 kyr between ~5.59 and 5.5 Ma. This interval contains two prominent glacial events (TG12 and TG14; Figure 9) that may have contributed to the final isolation of the Mediterranean. Although the progressive restriction of the Mediterranean was dominantly controlled by tectonic processes, even small glacioeustatic changes may have been important once the gateway became very shallow. The end of the intra-Messinian unconformity and resumption of deposition of the upper evaporite unit occurred at ~5.5 Ma [Krijgsman et al., 1999, 2001], or approximately eight precessional cycles below the Miocene/Pliocene boundary (5.33 Ma). This time corresponds to the transition between stages TG12 and TG11 and marks the end of the high-amplitude, glacial-interglacial cycles of the latest Miocene in both Site 982 and Morocco records (Figures 8 and 9). The resumption of sedimentation in Sorbas Basin following intra-Messinian desiccation is marked by the Zorreras Member, which consists of dominantly silts and sandy clays containing caspibrackish ostracod species that are characteristic of the Lago Mare fauna [Krijgsman et al., 2001].

Hodell et al. [1994] suggested erroneously that the decrease in $\delta^{18}O$ at the TG12/TG11 transition may correspond to the Miocene/Pliocene at 5.33 Ma on the basis of an extrapolation of 41-kyr $\delta^{18}O$ cycles from the top of C3An.1n. Because the age of the top of C3An.1n was taken to be 5.88 Ma [Shackleton et al., 1995a], as opposed to 6.04 Ma [Krijgsman et al., 1999a], the age of the TG12/TG11 transition was estimated to be ~160 kyr too young by Hodell et al. [1994]. The astronomically tuned chronology of Site 982 predicts an age of ~5.5 Ma for the TG12/TG11 transition and therefore this event predates the Miocene/Pliocene boundary by some 170 kyr.

After stage TG12, benthic $\delta^{18}O$ values were low from ~5.5 to 5.25 Ma and approximately equal to those prior to 6.3 Ma (Figure 9). Although benthic $\delta^{18}O$ values are relatively low at the Miocene/Pliocene boundary at 5.33 Ma, there is no sudden event that would indicate deglaciation and sea level rise as the cause of the reflooding of the Mediterranean. However, $\delta^{18}O$ values of bulk carbonate reach their minimum very close to the Miocene/Pliocene boundary at 5.33 Ma and then increase in two steps at 5.2 and 5.07 Ma (Figure 9). These early Pliocene increases in bulk carbonate $\delta^{18}O$ correlate to the top of the *Sphaeroidinellopsis* acme (5.20 Ma) and base of zone MPL2 (5.07 Ma), respectively, in the Mediterranean [Iaccarino et al., 1999]. Because benthic $\delta^{18}O$ is believed to be a more reliable indicator of global ice volume changes than is bulk carbonate, it is likely that major deglaciation and sea level rise predated the Miocene/Pliocene boundary (i.e., base of the Trubi Marls) by ~170 kyr (for an alternate view, see McKenzie et al. [1999]). Shackleton et al. [1995b] came to a similar conclusion and offered several explanations that might permit glacioeustatic sea level changes to play a role in terminating the MSC: (1) the age of the base of the Trubi Marls is actually 5.46 Ma because the lower cycles are related to obliquity and not precession; (2) sedimentation at the base of the Trubi did not resume until some time (~100 kyr) after the reflooding of the Mediterranean; (3) sea level rose at 5.46 Ma, but there was not enough erosion of the sill separating the Mediterranean and Atlantic to prevent desiccation until 5.32 Ma; (4) the oxygen isotope record of Site 846 is flawed and minimum values should occur at 5.32 Ma (stage TG5); or (5) the chronology of Site 846 is incorrect. Our results from Site 982 suggest that explanations 4 and 5 are not correct and 1 is unlikely because the link between sedimentary cycles in the Mediterranean and precession (not obliquity) has been confirmed by numerous studies (for a review, see Hilgen et al. [1999]). Explanations 2 and 3 remain viable options but are difficult to prove. For example, Kastens [1992] suggested that rising sea level allowed Atlantic waters to leak into the Mediterranean along east-west trending strike-slip faults. Down cutting of this narrow E-W trough would have eventually transformed it into a deep passageway until the Mediterranean had refilled to Atlantic sea level. This type of mechanism may account for the lag between sea level rise (as reflected in decreasing $\delta^{18}O$ values) and the resumption of normal marine sedimentation in the Mediterranean.

The correlations between Site 982 and Mediterranean sequences discussed above are based upon the cyclostratigraphy of Krijgsman et al. [1999, 2001]. We note, however, that there is considerable disagreement about the timing and nature of events, especially within the Messinian gap [Riding et al., 1998; 1999, 2000; Fortuin et al., 2000]. Riding et al. [1998, 1999] argued for a much older date for the intra-Messinian unconformity and drawdown of Mediterranean sea level than proposed by Krijgsman et al. [1999a, 2001]. They interpreted a hiatus between the Abad and Yesares Members, which correlates to an erosional unconformity separating the fringing reef unit from the terminal carbonate complex in marginal settings of the Sorbas Basin. The age of this unconformity is estimated to be between 5.9 and 5.5 Ma [Riding et al., 2000]. The authors proposed a sea level rise and subsequent reflooding of the basin beginning at 5.5 Ma, which resulted in evaporite deposition (Yesares Gypsum Member) in Sorbas as a barred basin. Normal marine conditions were soon established, giving rise to the deposition of marls, sands, and limestones of the Sorbas Member prior to the Miocene/Pliocene boundary. If this scenario is correct, then the reflooding of the basin beginning at 5.5 Ma coincided with the end of the late Miocene glacial period. Continued sea level rise in the latest Messinian is recorded in the Sorbas Member and is consistent with decreasing $\delta^{18}O$ prior to the Miocene/Pliocene boundary in Site 982. The occurrence of planktic foraminifera in latest Miocene sediments supports temporary marine incursion prior to the final inundation at 5.33 Ma [Iaccarino et al., 1999]. Although the timing and interpretation of events within the Messinian gap will continue to evolve, the astronomically tuned record of Site 982 provides a valuable template against which the timing of intra-Messinian events in the Mediterranean can be compared with processes occurring in the open ocean.

6. Conclusion

We developed an orbitally tuned timescale for the period from 7.5 to 4.5 Ma in ODP Site 982 on the Rockall Plateau in the North Atlantic Ocean. This interval spans the Messinian salinity crisis (MSC) and permits a direct correlation between the open-ocean Atlantic (Site 982), gateways connecting the Mediterranean and the Atlantic

(Rifian and Betic Straits), and the cyclostratigraphy of preevaporitic and evaporitic Messinian deposits in the Mediterranean. The first change in Mediterranean hydrography occurred near the Tortonian/Messinian boundary (~7.2 Ma) and is marked by an increase in benthic $\delta^{18}O$ values in the Salé Briqueterie Section of northwestern Morocco [Hodell et al., 1994]. No change is indicated in the benthic $\delta^{18}O$ record of Site 982 in the North Atlantic at this time, suggesting that the $\delta^{18}O$ increase in Moroccan sections was not related to an increase in global ice volume. Rather, the $\delta^{18}O$ increase represents cooling and/or increased salinity that resulted from circulation changes in the Rifian Corridor and the onset of a negative water budget in the Mediterranean [Hodell et al., 1994]. The restriction of the Mediterranean in the earliest Messinian was related to tectonic changes in the Betic-Rifian gateways rather than glacioeustatic sea level variations.

From 6.26 to 5.5 Ma a latest Miocene glaciation occurred and was marked by 18 glacial-to-interglacial oscillations that were dominantly controlled by the 41-kyr cycle of obliquity. Although the intensification of glaciation at 6.26 Ma may have contributed to the isolation of the Mediterranean, it preceded the base of the lower evaporite unit by some 300 kyr. The onset of evaporite deposition at 5.96 ±0.02 Ma [Krijgsman et al., 1999a] does not correspond to any major changes in the benthic $\delta^{18}O$ signal, supporting tectonic or local climatic forcing rather than a glacioeustatic cause for the inception of evaporite deposition [Weijermars, 1988; Martin and Braga, 1996; Krijgsman et al., 1999b, 2001].

The salinity crisis ended at 5.33 Ma with the restoration of open marine conditions as represented by the base of the Trubi Marls in Sicily. The end of the MSC occurred at a time of generally decreasing benthic $\delta^{18}O$ values, but we found no single benthic $\delta^{18}O$ low associated with the Miocene/Pliocene boundary. Oxygen isotopic values of bulk carbonate were lowest, however, at the base of the Pliocene, indicating strong interglacial conditions or changes in nannofossil assemblages. The most prominent decrease in benthic $\delta^{18}O$ occurred at 5.5 Ma with the transition from stage TG12 to TG11. This event preceded the Miocene/Pliocene boundary by 170 kyr, suggesting that sea level outside the Mediterranean began to rise prior to the base of the Pliocene. This implies that either a tectonic event was responsible for the end of the MSC or there was a significant lag between sea level rise and the resumption of normal marine sedimentation in the Mediterranean.

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