EUSTATIC SIGNALS IN DEEP-MARINE SEDIMENTARY SEQUENCES RECOVERED AT ODP SITE 978, ALBORAN BASIN, WESTERN MEDITERRANEAN SEA

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ABSTRACT: **A continuous section of Pliocene marine sediments was recovered at Ocean Drilling Program Site 978, located in the Alboran Sea between Spain and Morocco. Three Pliocene subunits have been defined at Site 978: the lowermost (Subunit IC, 129.2 m thick) is characterized by alternating beds of lighter, more calcareous, and darker, less calcareous, claystone with bioturbated upper and lower contacts (Type 1 cycles); the middle (Subunit IB, 67.1 m thick) is composed of relatively homogeneous nannofossil claystone; and the uppermost (Subunit IA, 211.6 m thick) contains abrupt-based darker, terrigenous layers interpreted as turbidites that are interstratified with lighter nannofossil claystone (Type 2 cycles). The rhythmically bedded light and dark layers in Subunit IC correlate with those in the Rosello Composite Section of Sicily, a global reference standard for the Pliocene time scale. These sedimentary cycles are products of variations in precession and resulting continental runoff. Missing cycles occur during eustatic highstands. The shift to more homogeneous sedimentation in Subunit IB is represented in similar-aged sequences throughout the Mediterranean which display evidence of submarine mass wasting. Mediterranean-wide slope degradation was likely a response to rapid sea-level change at approximately 3 Ma. This change in sedimentation style was accompanied by an upsection increase in sediment accumulation rates associated with turbidite influx in Subunit IA. Turbidite frequency throughout the Pliocene section can be linked to eustatic changes in sea level, with turbidite maxima corresponding with mid-sequence downlap surfaces and their associated condensed sections.**

INTRODUCTION

Perhaps the most studied Pliocene outcrops in the world are found in the central Mediterranean, particularly southern Sicily, where the type section is characterized by rhythmic bedding that has been linked to astronomical cycles. These units, the Trubi and overlying Narbone formations, have been the subject of detailed biostratigraphic, magnetostratigraphic, and cyclostratigraphic studies and form the basis for a global reference standard (Rosello Composite Section) for the Pliocene time scale (Cita and Gartner 1973; Rio et al. 1984; Hilgen 1987; Hilgen and Langereis 1989; Langereis and Hilgen 1991; Lourens et al. 1996; Hilgen et al. 1997). There has been some successful detailed correlation of cycles within these units with other onland marine sections across southern Italy and as far away as Crete, but previous attempts at correlating this section with sequences drilled by the Deep Sea Drilling Project (DSDP) and the Ocean Drilling Program (ODP) in other Mediterranean basins (e.g., ODP Site 653; Fig. 1) have been hampered by noncontinuous core recovery, poor magnetic properties, and a lack of distinct cyclic bedding (Langereis and Hilgen 1991). Such correlation is now possible with the discovery of cyclically bedded Pliocene units in a submarine section at ODP Site 978, over 1,400 km west of Sicily, in the Eastern Alboran Sea (Fig. 1; Comas et al. 1996).

The stratigraphy of Site 978 is detailed in Comas et al. (1996) and summarized here in Table 1. Bedding couplets in the lower Pliocene section at Site 978 (Fig. 2, Subunit IC) consist of alternating carbonate-poor and carbonate-rich intervals that are unlithified to semilithified equivalents of the marl–limestone couplets observed elsewhere in the rock record and attributed to orbital-climatic variations and atmosphere–ocean feedback systems (e.g., Einsele and Ricken 1991). These bedding couplets were referred to as Type 2 cycles by shipboard scientists (Comas et al. 1996). Subunit IC is overlain by a homogeneous hemipelagic unit (Subunit IB), which we believe roughly corresponds with a geologically ''sudden'' climatic step or event at approximately 3 Ma resulting from the onset of Northern Hemisphere glaciation and Pleistocene-type climatic fluctuations (see Berger 1982; Backman and Pestiaux 1986; Raymo et al. 1986; Einarsson and Albertson 1988; Zachariasse et al. 1989). Note that the distribution of ice-rafted detritus suggests that this onset was stepwise between 3.5 and 2.4 Ma, with a major expansion of the Greenland ice sheet at 3.3 Ma spreading to Scandinavia and North America at around 2.72–2.75 Ma (Flesche et al. 2002). The homogeneous interval is succeeded by an upper Pliocene section (Subunit IA) that contains coarse clastic interbeds, which also appear to be rhythmic (referred to as Type 1 cycles).

Herein we discuss the nature and origins of two styles of rhythmic bedding found in the upper and lower Pliocene sections at Site 978, as well as of an intervening homogeneous unit within the middle Pliocene. We then summarize the nature of Pliocene deep marine sedimentation across the Mediterranean using DSDP and ODP results, as well as descriptions of uplifted Pliocene sediments in Sicily, Calabria, and Crete. Our synthesis and correlation between onshore Pliocene sections and the Pliocene section at Site 978 suggest that eustasy significantly influenced deposition of Pliocene deep-marine sequences in the Alboran Basin during a period of regional compressional tectonics. Thus, the Alboran example shows that eustatic signals may be recorded in deep-marine (distal) sequences within tectonically active basins, in contrast to the passive-margin basins, where these signals are perhaps best defined.

PLIOCENE SECTION AT SITE 978

The Alboran Basin has had a complex tectonic history, having formed during Miocene extension and since been subjected to compressional tectonics (Comas et al. 1999). Site 978 is situated in a small east–west trending subbasin approximately 60 km south of Cabo de Gata along the southern coast of Spain (Fig. 1). The hemipelagic Pliocene section in the Alboran Basin was deposited in water depths up to 1500 m (Stanley 1977) after post-Messinian flooding and the establishment of normal marine conditions in the Mediterranean Sea.

At Site 978, a thick (400 m) Pliocene section was rotary drilled with almost complete core recovery except near the Pleistocene and Miocene contact zones (see recovery column in Fig. 2 and accumulation curve in Fig. 3). Detailed shipboard analyses (Comas et al. 1996) provide infor-

JOURNAL OF SEDIMENTARY RESEARCH, VOL. 74, NO. 3, MAY, 2004, P. 379–391
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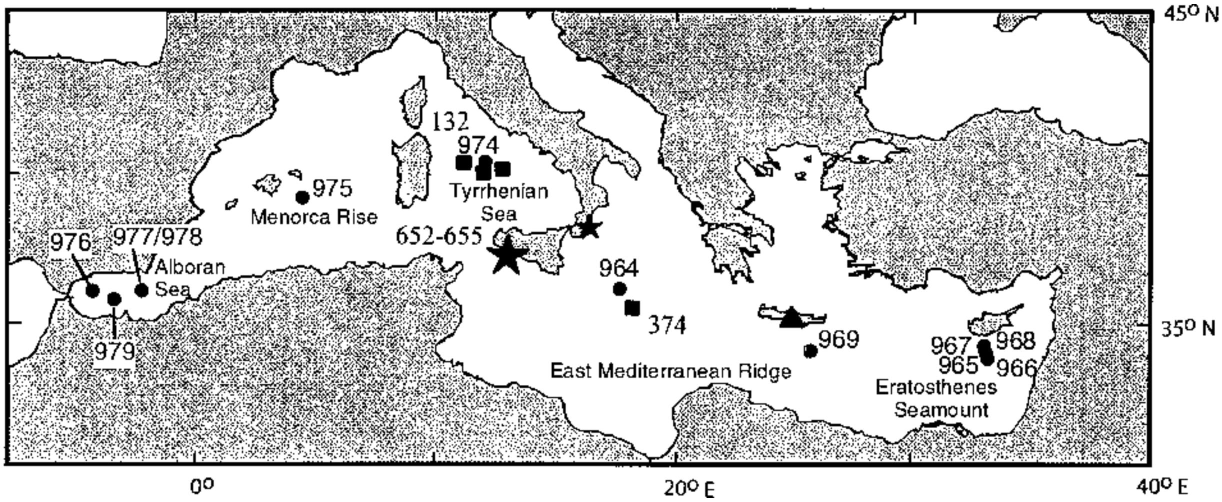


FIG. 1.—Map of the Mediterranean Sea showing locations of ODP Leg 161 sites and other ODP and DSDP sites referred to in the text. Leg 160 and 161 (964–979) site locations are marked by filled circles and other sites by filled squares. Approximate locations of outcrops of Trubi and Narbone formations are indicated by large (Sicily) and small (Calabria) stars. Equivalent deposits in Crete are marked by a triangle.

mation on sedimentary structures, sediment composition, and sediment geochemistry throughout this interval. These data allowed shipboard scientists to subdivide the Pliocene into three subunits as described in Table 1. The most striking features in Subunits 1A and 1C are alternating light and dark cycles (rhythmic bedding). The two types of cycles are described in Table 1 and illustrated in Figure 4. Their downhole distributions are summarized in Figure 5. The subunit boundaries and cyclicity within subunits are apparent in the down-hole carbonate distribution shown in Figure 6.

Note that for descriptive purposes we have adopted the tripartite subdivision of the Pliocene (upper, middle, and lower) proposed by Cita et al. (1996) because it loosely fits the three subunits defined at Site 978. The middle Pliocene roughly corresponds to the Gauss magnetic polarity chron and foraminiferal stages MPL4b and MPL5a with an approximate age span from 2.6 to 3.55 Ma (Berggren et al. 1995; Cita et al. 1996).

TABLE 1.—*Site 978 lithostratigraphy.*

Subunit IA
Interval—213.0 to 342.2 meters below the sea floor (mbsf).
Age—late Pliocene
Major Lithology—olive gray to dark greenish gray nannofossil clay.
Minor Lithologies—sand, silt, silty clay, and clay that constitute slightly darker, less calcareous interbeds that range in total thickness up to 65 cm (see Fig. 4, p. 361, and Fig. 8, p. 364, in Comas et al. 1996). These begin at abrupt bases and are commonly overlain by medium-dark gray sand to silt intervals that range up to 10 cm in thickness and locally exhibit normal grading and parallel lamination. Sandy or silty intervals pass up into lighter (dark greenish gray to grayish olive), bioturbated (*Planolites, Chondrites*, and *Zoophycos*) nannofossil-rich to calcareous, silty clay to clay. These dark intervals were designated by shipboard scientists as Type 1 beds and in combination with overlying lighter layers as Type 1 cycles.

Subunit IB
Interval—342.2 to 409.3 mbsf
Age—middle Pliocene
Major Lithology—fairly homogeneous nannofossil clay with minor bioturbation.
Minor Lithologies—few, poorly developed, darker beds similar to those found in Subunits IA and IC.

Subunit IC
Interval—409.3 to 620.9 mbsf.
Age—early Pliocene
Major Lithologies—alternating light (greenish gray to light olive gray to dusky yellow green) and dark (dark greenish gray to grayish olive to olive gray) beds of nannofossil or calcareous claystone and nannofossil or calcareous silty claystone. Lighter layers contain abundant *Planolites* burrows, and the dark layers *Chondrites* burrows. The majority of dark intervals in Subunit IC exhibit bioturbated to gradational upper and lower contacts (see Fig. 11, p. 365, and Fig. 12, p. 366, in Comas et al. 1996). At burrowed contacts, burrows are commonly filled with sediment from the overlying bed. Shipboard scientists designated these dark intervals as Type 2 beds and in combination with overlying light layers as Type 2 cycles. In general, the Type 2 beds are thicker and more prevalent in the section below 380 m. Some Type 1 beds are also present in Subunit IC. This subunit is more indurated than overlying subunits.

PLIOCENE SECTION AT OTHER (ODP/DSDP) DRILL SITES

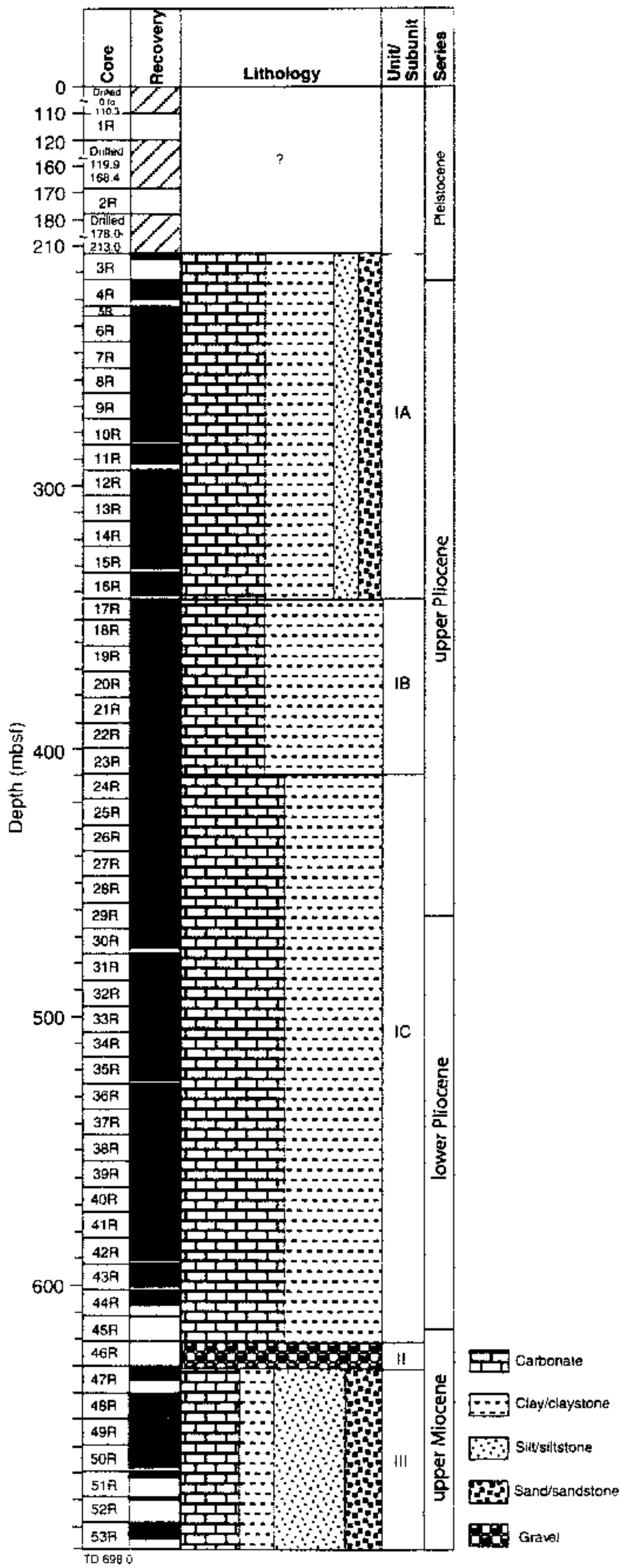
Although sediments of Pliocene age have been recovered at many Mediterranean DSDP sites, the practice of spot coring, poor recovery rates, and drilling deformation have resulted in only a few sites where relatively continuous Pliocene sections were recovered. Drilling techniques and strategies have improved and more complete sections have been recovered during recent ODP cruises. We examined core photographs and shipboard descriptions from all the Pliocene-age cores recovered at Mediterranean DSDP and ODP sites to document the occurrence of cyclic sedimentation similar to that observed at Site 978, as well as the nature of sedimentary successions of middle Pliocene age (~3 Ma). The results of this compilation, including some reinterpretation of core structures, are summarized in Table 2 (see Fig. 1 for site locations).

Cyclic light–dark layers are present in Pliocene sections at several DSDP and ODP sites in the Atlantic Ocean: for example, at Site 398 on the Vigo seamount (Maldonado 1979), and at Sites 532, 1081, 1082, 1084, 1085, 1086, and 1087 on the Walvis Ridge (Dean and Gardner 1985; Diester-Haass and Rothe 1988; Wefer et al. 1998). Recently, such cycles were also identified on electric logs of marine sequences drilled at several petroleum industry wells in the Atlantic Ocean (Gulf of Cadiz) just west of the Strait of Gibralter (Sierro et al. 2000).

PLIOCENE OUTCROPS IN ITALY AND CRETE

The Pliocene of southwestern Sicily (Fig. 1) is subdivided into the Trubi Formation and the overlying Narbone Formation. The Trubi Formation is composed of fossiliferous, thick- (1 m) to medium-bedded, intensely bioturbated, semi-indurated pelagic sediments characterized by quadripartite (Fig. 7), small-scale sedimentary cycles consisting of gray and beige chalk to marl alternating with white chalk (Cita and Gartner 1973; Hilgen 1987; Langereis and Hilgen 1991). The carbonate fraction consists predominantly of poorly preserved calcareous nannofossils (Rio et al. 1984) and planktic foraminifers with minor benthic foraminifers and ostracodes (Cita and Gartner 1973). Carbonate content varies from 60–80% in the lower half of the Trubi Formation to 50–70% in the upper half (Cita and Gartner 1973; de Visser et al. 1989; Hilgen and Langereis 1989).

The top of the Trubi Formation locally contains disturbed (slumped) intervals (Hilgen 1987; Langereis and Hilgen 1991). The contact between the Trubi and Narbone formations is gradational to locally unconformable



where it is marked by a possible slump scar (Rio et al. 1984; Langereis and Hilgen 1991). The age of the contact is approximately 3.0 Ma, roughly coincident with the M subchron of the Gauss chron (Rio et al. 1984; Langereis and Hilgen 1991). The younger Narbone Formation is composed of silty marl intercalated with brown, organic-rich layers (sapropels) (Rio et al. 1984; Lourens et al. 1996). It is more enriched in terrigenous components than the Trubi Formation, and this accounts for the pronounced increase in sediment accumulation rate from approximately 70 m/My in the Trubi Formation to 160 m/My in the Narbone Formation (Cita and Gartner 1973; Rio et al. 1984).

The Trubi-equivalent chalks and marls in Calabria (Hilgen 1987) and Crete (Jonkers 1984) are also rhythmically bedded, but with marl doublets rather than the quadruplets of southern Sicily. Specifically the Calabria rhythms lack a beige interbed and consist of alternating white (or beige) carbonate-rich and gray beds (Fig. 7). In the Crotone–Spartivento Basin of southern Calabria (Fig. 1), units similar to the Sicilian Trubi and Narbone formations are separated by a major hiatus (Hilgen 1987) which spans approximately 3.75 to 3.0 Ma.

Milankovitch-scale cycles have also been recognized in fluvial and lacustrine sequences of Pliocene age in the eastern Mediterranean Region (Abreu et al. 2000; Nummedal 1999; van Vugt et al. 1998; van Vugt et al. 2001). Precessional cycles like those that characterize the Pliocene at Site 978 are most distinctly expressed in carbonate lacustrine sequences (van Vugt et al. 2001).

ORIGIN OF LOWER PLIOCENCE CYCLES IN THE MEDITERRANEAN REGION

During the Pliocene, water depths ranged from 1000 to 1900 m across the western and central Mediterranean. As a consequence, sediments at Site 978 and the other outcrop localities were likely deposited above the lysocline during this period (Wright 1978; Comas et al. 1996). Above the lysocline, light–dark cycles can be produced by variations in carbonate input, clay input, or carbonate dissolution (e.g., Diester-Haass 1991; Einsele and Ricken 1991; Schwazacher 1993). Where carbonate dissolution or carbonate productivity is thought to be the dominant cause of sediment cyclicity, the limestone beds are thicker than the more clay-rich beds (Einsele and Ricken 1991). Although clay-rich beds in the Trubi and Narbone formations are thinner than their associated carbonate-rich intervals, carbonate dissolution and/or decreased carbonate productivity have not been favored as mechanisms for producing the carbonate-poor gray layers because benthic to planktic foraminifer ratios are relatively constant and dissolution-prone taxa are usually well preserved (e.g., Spaak 1983; Hilgen 1987; Diester-Haass 1991). Furthermore, carbon and oxygen isotopic analyses of foraminifers in Pliocene rhythmic units in Crete and Sicily suggested that the rhythms are a product of alternating relatively dry and wet intervals, corresponding to carbonate-rich and carbonate-poor periods, respectively (Jonkers 1984; Gudjonsson and van der Zwann 1985; Gudjonsson 1987; de Visser et al. 1989; Lourens et al. 1992; Van Os et al. 1994). The increased runoff during wet periods was also likely associated with an increase in terrigenous clastic input (e.g., Spaak 1983; Jonkers 1984; Gudjonsson and van der Zwaan 1985; Hilgen 1987). Thus the favored mechanism for producing the gray layers is by terrigenous dilution.

A comparison and detailed bed-by-bed correlation (see below) between

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FIG. 2.—Simplified stratigraphic column for Site 978 taken from Comas et al. (1996). Note that most of the Pleistocene section (to 213 mbsf) was not cored at this site in order to maximize drilling time in the Upper Miocene section. Black intervals in recovery column represent recovered core. Note the following errors in the preliminary results volume (Comas et al. 1996): the core photographs for 33R and 34R (p. 864–865) should be switched; and the smear-slide data (Section 4, p. 949–987) are missing depths, so that the component headings should be shifted over one column.

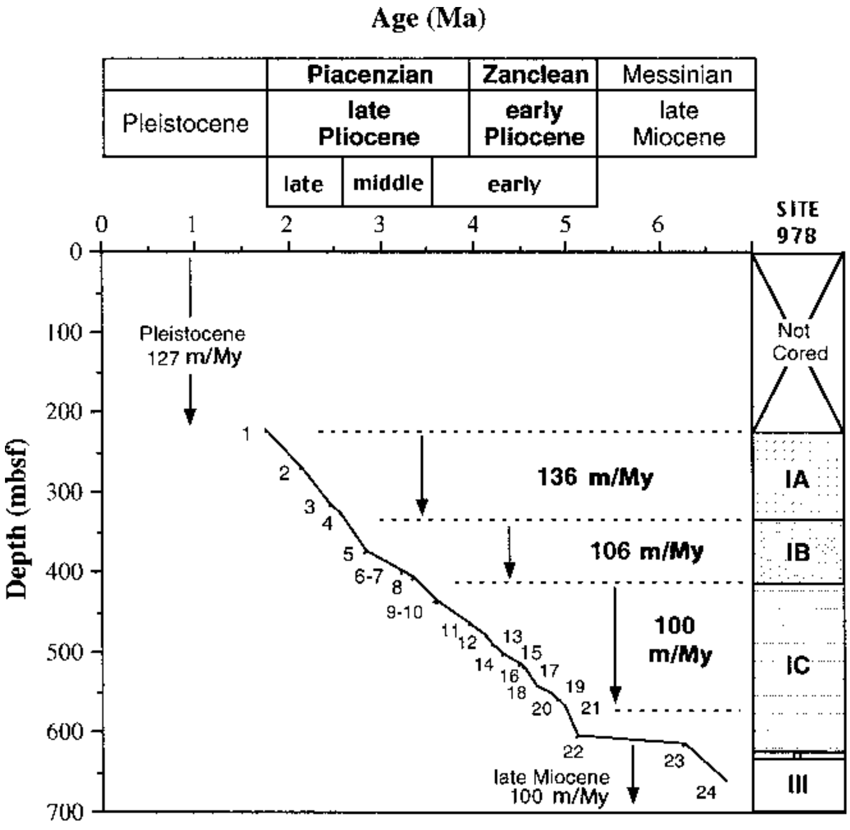

FIG. 3.—Plot of age vs. depth for Site 978 after Comas et al. (1996). Numbers on curve refer to biostratigraphic events and magnetostratigraphic data.

individual beds within the Trubi Formation and those in Unit IC at Site 978 suggests that Type 2 rhythms of the latter are the products of similar processes. The Type 2 muddy interbeds at Site 978 have gradational lower contacts, are structureless in that they lack internal lamination or grading, and are bioturbated throughout. These characteristics appear to preclude a mud turbidite interpretation (see Piper and Stow 1991 and Stanley 1985) and suggest instead that they were produced by a gradual process, such as an increase in suspension fallout of fine-grained terrigenous material and organic matter, or perhaps changes in carbonate dissolution or production rates.

Internal zoning within the lighter carbonate-rich intervals, such as the white–beige–white pattern found in the Trubi Formation, is not readily apparent at Site 978. The unique clay mineralogy (palygorskite and kaolinite) and larger grain size of the beige layers in the Sicilian Trubi were attributed by de Visser et al. (1989) and Foucault and Mélières (2000) to increased eolian influx from North Africa. Thus there may have been minimal or no eolian input from Africa at Site 978, similar to the Calabrian and Cretan sections. Alternatively, the signal may be present but not as easily discerned in unweathered core at Site 978 as in weathered outcrops.

Additional information on the environment of deposition of the Site 978 cycles can be gleaned from bioturbation patterns. Outcrops of Trubi marls have been described as bioturbated, but without detailed trace-fossil description and interpretation. In contrast, trace fossils were easily discerned and identified in freshly slabbed cores at Site 978. The lighter carbonate-rich layers tend to be characterized by abundant *Planolites* burrows, and the darker terrigenous layers by *Chondrites* (see Comas et al. 1996). These changes in trace-fossil assemblages could be a function of changing oxygen concentration of interstitial water (Ekdale and Mason 1988), with *Chondrites* tolerating slightly lower oxygen levels (Bromley and Ekdale 1984; Ekdale et al. 1984; Ekdale 1985). In addition grain size may be an important factor, because *Planolites*-producing organisms prefer coarser sediment (Wetzel 1991). Thus the tendency for lighter, more calcareous layers to exhibit *Planolites* burrows may be tied to the higher content of sand-size foraminifers in these beds.

Diagenesis must also be considered in the interpretation of these distinctive cycles. In the lower Pliocene section at Site 978 (Subunit IC), the difference in carbonate content between adjacent light and dark beds is generally about 15% (Fig. 6). This is similar to differences seen in other rhythmic sequences (e.g., Quaternary to late Pliocene, DSDP Sites 532 and 362; Diester-Haas and Rothe 1988; Diester-Haass 1991). However, this difference increases slightly to about 20% down section (Fig. 6) and is evident in the pronounced downhole increase in color contrast between light and dark layers. Smear-slide analyses of the more carbonate rich intervals at the base of the rhythmic Pliocene section indicate that these intervals contain higher percentages of nannofossils than the carbonate-rich beds higher in Subunit IC. This could indicate either that the foraminiferal component has been preferentially dissolved or that nannofossil accumulation rates were higher during the early Pliocene at this site. In fact, foraminifers in the lower Pliocene section (Cores 978A-33R-CC to 45R-CC) do show evidence of enhanced dissolution, fragmentation, and deformation consistent with increased compaction and pressure solution (Comas et al. 1996). Furthermore, the geochemistry of organic compounds present in this lower section suggests that they have experienced a thermal event or pulse (Comas et al. 1996) that could have enhanced the loss of carbonate from the shales and precipitation of carbonate in the carbonate-rich intervals as nannofossil overgrowths and/or micrite. Pore-water compositions in Subunit IC suggest influence by a paleobrine, possibly filtering from the underlying gravelly Unit II up into the sediments of Unit I. Furthermore, alkalinity, Mg concentration, and Sr concentration indicate a pulse of carbonate precipitation and recrystallization (Comas et al. 1996). Enhanced acoustic velocities below approximately 550 mbsf at the base of Subunit IC may also reflect increasing lithification and/or recrystallization and/or cementation (see Comas et al. 1996).

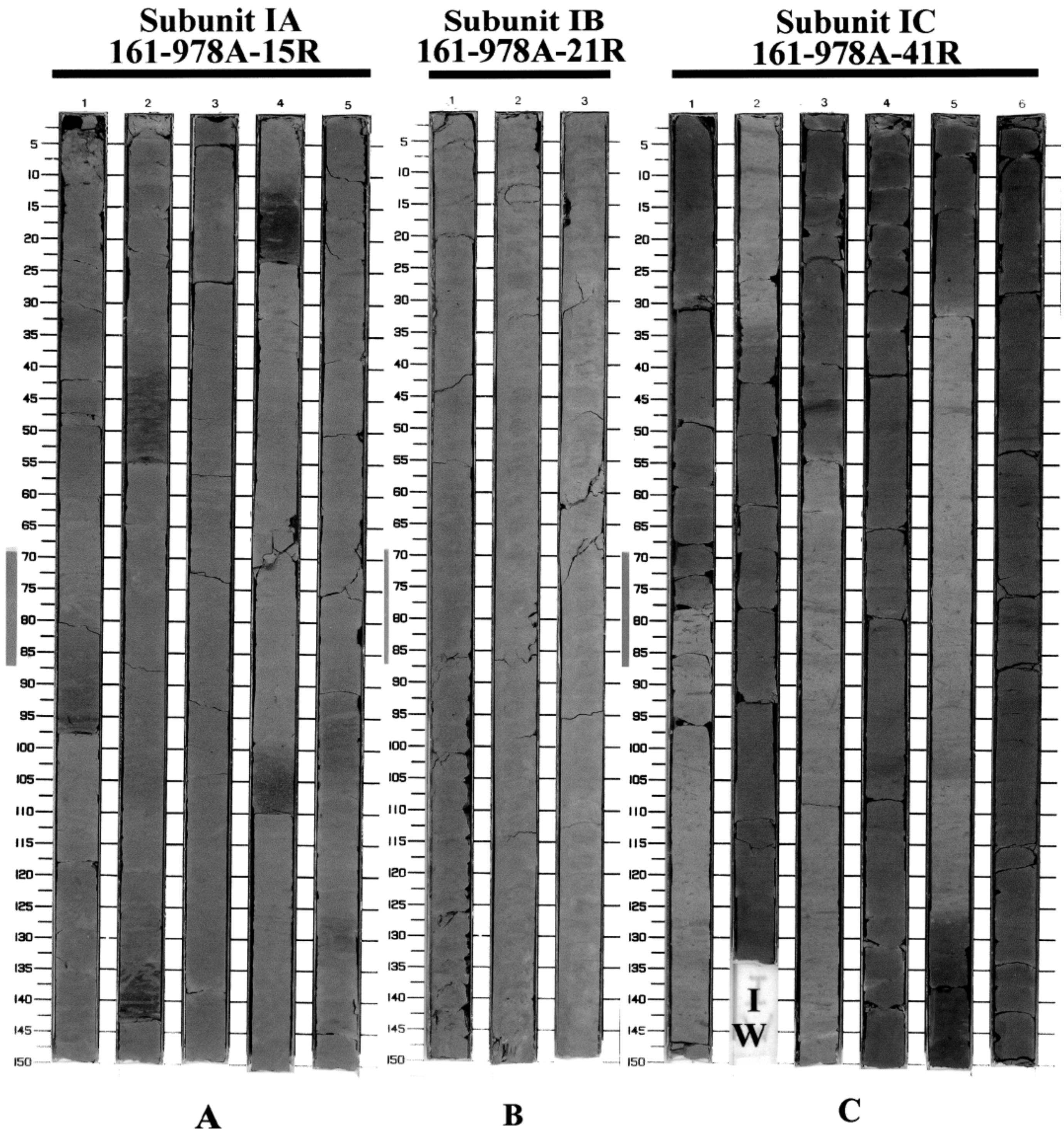

FIG. 4.—Representative whole-core photographs for subunits at Site 978. **A)** Photograph of Core 161-978A-15R in Subunit IA. Scale is in centimeters, and the top of the core is at 322.8 mbsf. Distinct Type 1 dark beds are present in Sections 1 (95 cm), 2 (50 and 140 cm), and 4 (15 and 105 cm). Each exhibits a sharp base, typically overlain by faintly laminated sandy silty clay that passes upward into strongly bioturbated nannofossil clay. Bioturbated, less distinct Type 1 dark beds are also present in Section 5 (110 and 130 cm). **B)** Photograph of Core 161-978A-21R in Subunit IB, showing featureless homogeneous beds that characterize this subunit. Scale is in centimeters and the top of the core is 380.6 mbsf. **C)** Photograph of Core 161-978A-41R in Subunit IC. Scale is in centimeters, and the top of the core is at 572.8 mbsf. Distinct, decimeter-scale beds of lighter (more calcareous) and darker (Type 2; less calcareous) nannofossil claystone alternate throughout the core. IW refers to a section of the core removed for interstitial water sample.

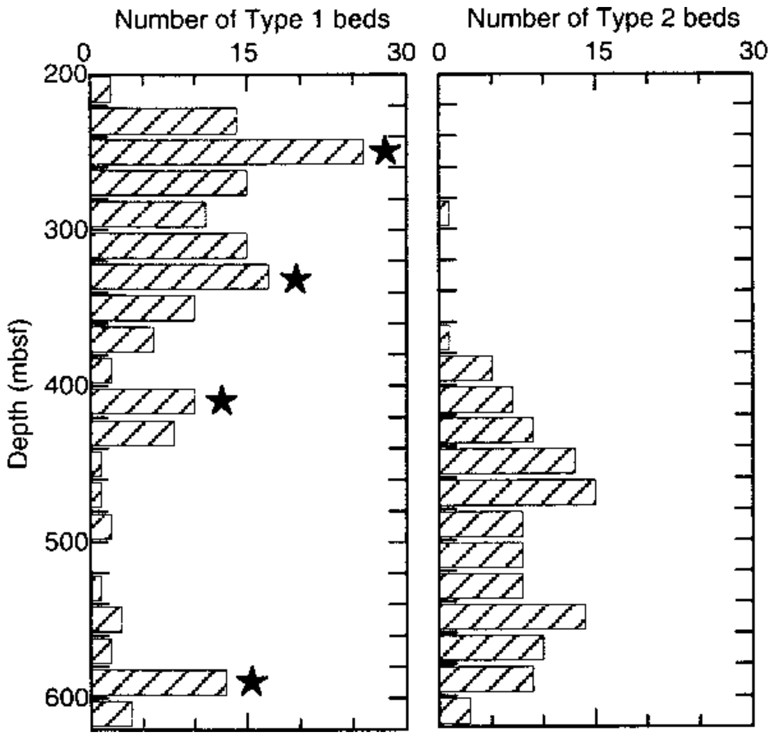

FIG. 5.—Histogram plots of numbers of Type 1 and Type 2 dark beds at Site 978 versus depth. Stars denote maxima in numbers of Type 1 dark beds. See text for discussion. Modified from Comas et al. (1996).

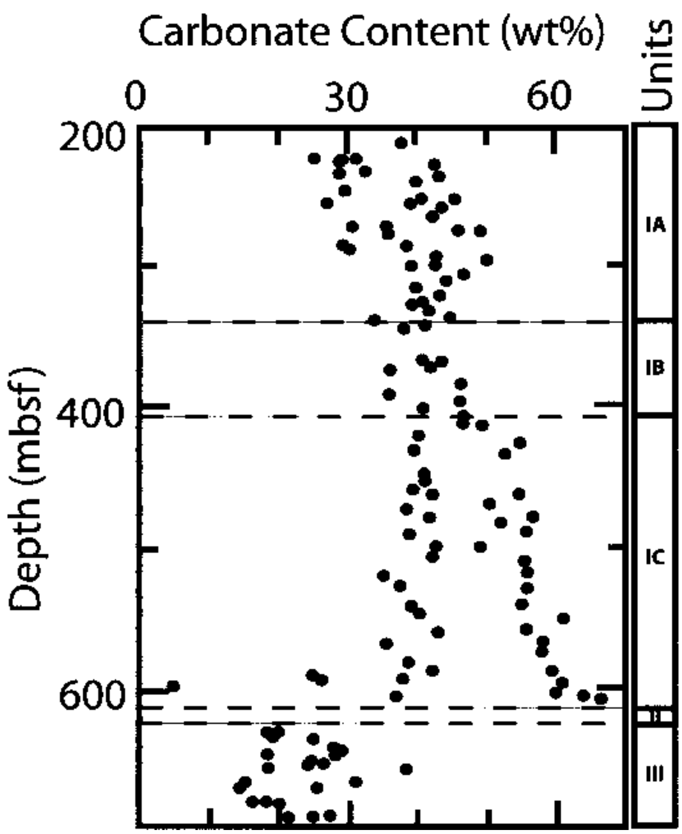

FIG. 6.—Plot of carbonate content at Site 978 versus depth. Carbonate content was determined analytically; data are given in Comas et al. (1996). Dashed lines mark Unit and Subunit boundaries.

DETAILED CORRELATION OF THE LOWER PLIOCENE SECTION AT SITE 978 WITH THE TRUBI FORMATION USING BIOSTRATIGRAPHY, MAGNETOSTRATIGRAPHY, AND CYCLOSTRATIGRAPHY

Rhythmic marl–limestone sequences have been widely attributed to changes in ocean–atmosphere feedback systems that amplify orbital variations such as Milankovich cycles associated with variations in precession that exhibit major periods of 19 and 23 ky and an average quasi-periodicity of 21.7 ka (Berger and Tricot 1986). Indeed, Hilgen (1987), Lourens et al. (1996), and Zachariasse et al. (1989) interpreted the cycles in the Sicilian sections to result from precessional variation with an average duration of 21.7 ka per cycle. They suggested that the gray and beige marls correlate, respectively, with the amplitudes of the summer insolation maxima and minima. According to Zachariasse et al. (1989), the gray carbonate-poor intervals are associated with higher seasonality reflected in higher summer and/or lower winter surface water temperatures. In addition, Lourens et al. (1996) cited the presence of thicker composite lighter carbonate-rich cycles as evidence for the influence of obliquity.

Hilgen (1987) correlated the Calabrian white carbonate-rich beds with the Sicilian white–beige–white intervals and considered the gray beds in both sections as time equivalent (Fig. 7). He found essentially the same number of basic rhythmites (white–gray and white–beige–white–gray bedding sequences) in both sections and a correspondence of thicker (composite) rhythmite intervals. The calculated mean duration of the Calabrian Trubi-equivalent rhythms is 19.9 ky (Hilgen 1987). A similar bed-by-bed correlation (even of thicker, composite beds) can be made with Pliocene rhythmically bedded units in Crete, but only in the intervals deposited from 3.03 to 3.65 Ma (Hilgen 1987).

The calculated recurrence interval for the Type 2 beds in Subunit IC at Site 978 is approximately 25 ky, similar to frequencies cited above for the Trubi Formation and equivalents. We compiled a log of the light–dark couplets in Subunit IC at Site 978 based on core photographs and descriptions presented in Comas et al. (1996). Assuming that the more complex, quadripartite cycles of the Trubi Formation (gray–white–beige–white) correspond to the simple dark–light couplets observed at Site 978 (Fig. 7), a general

TABLE 2.—*Pliocene Record at Mediterranean DSDP/ODP Sites.*

Site	Location	Comments/Interpretation*	Published Source(s)
Eastern Mediterranean			
Site 965	Top of Eratosthenes Seamount	Middle Pliocene (~ 3 Ma) hiatus (slump scar)	(Emeis et al. 1996)
Site 966, 967, 968	Flank of Eratosthenes Seamount	Sapropels (organic-rich layers) in upper Pliocene; middle Pliocene sections (below and above 3 Ma horizon) show evidence of mass movement including debris-flow deposits.	(Emeis et al. 1996)
Site 969	Near Crete	Sapropels (organic-rich layers) in upper Pliocene; middle Pliocene sequence includes a 25-m-thick homogeneous layer with some evidence of mass movement/soft-sediment deformation.	(Emeis et al. 1996)
Site 964	Ionian Basin	Sapropels (organic-rich layers) in upper Pliocene; middle Pliocene section consists of structureless nannofossil–foraminiferal ooze without stratification.	(Emeis et al. 1996)
Site 374	Ionian Basin	3-Ma-old sediments occur at approximately 340 mbsf where there is some drilling (?) disturbance and inclined bedding planes that could possibly be a product of mass movement.	(Hsü et al. 1978)
Western Mediterranean			
Sites 652–655	Tyrrhenian Sea	Sapropels (Organic rich layers) in upper Pliocene; some dipping bedding planes at approximately the 3 Ma horizon may be a product of mass movement or drilling deformation; no distinct cycles in lower Pliocene.	(Kastens et al. 1987)
Site 974	Tyrrhenian Sea	Sapropels (organic-rich layers) in upper Pliocene; 3 Ma horizon lies at approximately 125 mbsf in a section of nannofossil clay with minor color banding and bioturbation, above and below which the Pliocene sediments have been affected by soft-sediment deformation and slumping; no distinct cycles in lower Pliocene.	(Comas et al. 1996)
Site 132	Tyrrhenian Rise	Continuous Pliocene section; negligible terrigenous input; bioturbated, no stratification.	(Ryan et al. 1973)
Site 975	Balearic Basin	Thin (< 10 cm) dark intervals in lower Pliocene; 3 Ma horizon lies at approximately 200 mbsf, above which the latest Pliocene sediments (115 mbsf) have been affected by soft-sediment deformation and slumping.	(Comas et al. 1996)
Site 976	Alboran Sea	Color banding in lower Pliocene; 3 Ma horizon lies between 500 and 550 mbsf, above which the sedimentary section consists of a poorly recovered series of sandy turbidites.	(Comas et al. 1996)
Site 977	Alboran Sea (28 km from Site 978)	Below the 3 Ma horizon, there is evidence of mass movement (soft-sediment folding, intraclastic breccia/conglomerate); major hiatus (1 My) may represent a basal slide plane; Lower Pliocene contains rhythmic bedding, light–dark couplets similar to those at Site 978, but disturbed by drilling (intensely biscuited).	(Comas et al. 1996)

Note: Sand beds and laminae (turbidites) similar to the Type 1 beds described at Site 978 were noted in a number of Pliocene cores (e.g., Sites 121–124, 371; Ryan et al. 1973 and Hsü et al. 1978), but the core photographs were of poor quality and recoveries were spotty or low across these intervals, so they were not included as ''cycles''.

* In many instances our determination of mass-movement, slump, and debris-flow deposits is based on reinterpretation of core photographs.

correlation can be made between the two sections (Fig. 8). For correlation purposes, we have used the Rosello Composite Section, which integrates paleomagnetic, biostratigraphic, and cyclostratigraphic data for a number of sections of Sicilian Trubi Formation (Langereis and Hilgen 1991).

The correlation between the Trubi Formation and Subunit IC at Site 978 is facilitated by the excellent recovery and minimal core deformation (biscuiting) at Site 978, as well as the establishment of well-defined biostratigraphic and magnetostratigraphic datums for both the Trubi Formation (e.g., Hilgen and Langereis 1989) and Subunit IC at Site 978 (Comas et al. 1996). Specific datums used in this correlation include top of MPL1, top of C3n.3n (S), top of MPL2, top of C3n.1n (C), top of MPL3, top of MPL4a, and top of C2An.3n (within MPL4b). Above the last datum (approximately 410 mbsf at Site 978), cycles are poorly defined and no attempt was made at correlation. Trubi beds nearest these datums are considered ''key'' beds and are numbered in Figure 8 according to the scheme outlined in previous correlation studies (e.g., Hilgen and Langereis 1989). For simplicity, we refer in our discussion to intervals by these numbered key beds rather than by the datums. On the basis of the ages of datums, the time represented by these intervals between these key beds ranges from 0.24 to 0.48 My. In one instance at Site 978 (arrow in Fig. 8), it is likely that a carbonate-rich bed fell completely into an unrecovered zone; to account for this we have added a cycle to the Site 978 total (11) between key beds 60 and 44. As defined in previous studies of the Trubi Formation, there are a number of amalgamated light intervals (white–beige–white) in the Trubi Formation that lack intervening gray layers. These anomalously thick units are indicated by stars, with the number of stars referring to the number of

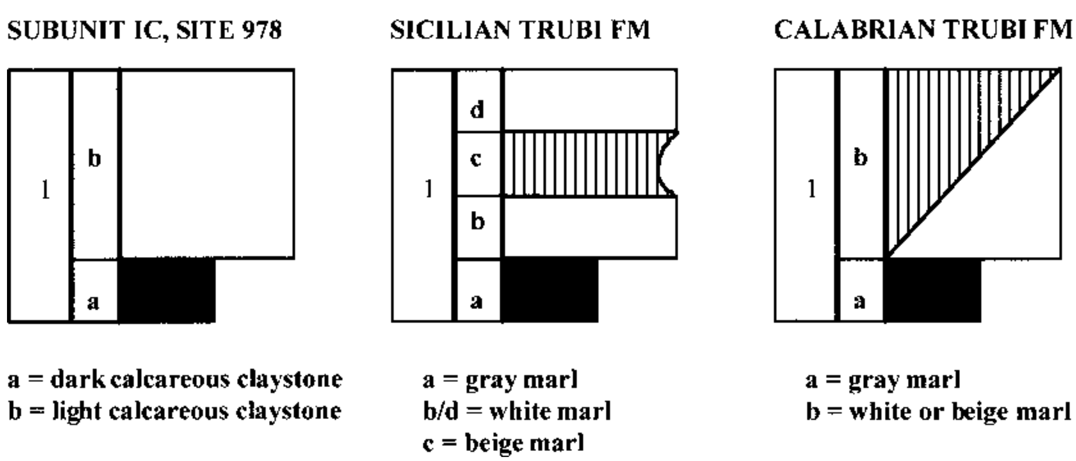

FIG. 7.—Schematic outline of lithologies that constitute equivalent cycles in Sicilian and Calabrian Trubi Formation (modified from Hilgen 1987) and Subunit IC, Site 978. The Trubi examples represent relative weathering and induration profiles with white and beige marls more indurated than gray marls.

rhythms represented by each. We indicate in the Site 978 stratigraphic column where we believe such amalgamation may exist using the same star system. Finally, the difference in scales between the two sections in Figure 8 should be noted; the sediment accumulation rate of Subunit IC sediment at Site 978 (200 m thick) was approximately twice that of the equivalent Trubi Formation (90 m thick).

Several intervals of thick light-colored beds in the Rosello section (e.g., intervals 5–6, 20–22, and 38–40 in Fig. 8), thought to be associated with the effects of obliquity, appear to correspond roughly with intervals of thicker (more amalgamated?) light beds at Site 978. In all of the Site 978 intervals in Figure 8, the number of cycles is the same or less than in correlative intervals of the Rosello Composite Section, implying that the Rosello composite section is more complete. The implications of the incomplete intervals with missing cycles are discussed below.

CYCLE ORIGIN IN SUBUNIT IA AT SITE 978

The uphole change from dominance of Type 2 beds in Subunit IC to Type 1 beds in Subunit IA occurs at approximately 380–400 mbsf at Site 978, indicating a major change in sedimentation style at approximately 3 Ma. On the basis of sediment accumulation curve (Fig. 3), sediment accumulation rates also change at this point, increasing from approximately 100 m/My below 400 mbsf to 140 m/My above 400 mbsf. We believe this increase is best explained by a post-3 Ma increase of turbidity-current-derived terrigenous material. The change from predominantly Type 2 to Type 1 beds is shown in the frequency plots (number of beds per 20 m interval) in Figure 5. Note that there are some Type 1 beds within the lower Pliocene section (Subunit IC) and that frequency maxima of Type 1 beds increase in magnitude upsection in Subunits IB and IA.

The sedimentary structures associated with Type 1 dark beds, namely abrupt basal contacts, laminated to cross-laminated sandy or silty bases, and bioturbated tops, suggest that they were likely deposited by low-density turbidity currents of the type described by Pickering et al. (1989). The calculated recurrence interval for these beds in the upper Pliocene section is approximately 9,400 yr. Their periodicity, roughly 10,000 yr, is consistent with sixth-order glacio-eustatic changes for turbidites in other sedimentary sequences (Einsele et al. 1991; Vail et al. 1991).

DISCUSSION OF EUSTATIC SIGNALS IN THE PLIOCENE SECTION AT SITE 978

Missing Rhythms (Milankovitch Cycles) in Subunit IC (Lower Pliocene)

The intervals in Subunit IC at Site 978 that contain less-than-expected numbers of cycles occur between key beds 35–44 (4.18–4.48 Ma) and 60–78 (3.57–3.94 Ma). In these two intervals, the Site 978 section is thinner than would be expected if sediment accumulation rates had been uniform. Variations in bed thickness in the 60–78 interval (Fig. 8) suggest that the problem may be concentrated in the lower part of this interval.

There are a number of possible causes for the mismatches in numbers of cycles. Some lighter cycles may have subtle signatures that were missed during our shipboard description. Other cycles may be amalgamated or obscured by bioturbation. In addition, section could be missing because of minor normal faulting, removal by slumping, removal by erosive bottom currents, or hiatuses caused by periods of nondeposition. Faulting is unlikely because bedding is essentially horizontal across these zones and no normal faults were identified in the cores (Comas et al. 1996) or on a seismic section across Site 978 (Tandon et al. 1997). No evidence of mass movement (e.g., slump folds or intraclastic breccias) was observed within these two intervals; however, thin (1–2 cm), parallel-laminated foraminiferal and bioclastic sandy to silty layers present throughout Unit I are probably the winnowed products of contour currents. This suggests that bottom currents may have been particularly strong in the vicinity of Site 978 during these time intervals, resulting in removal or nondeposition of some section.

In at least one of these intervals there is further evidence of erosion or nondeposition. A reexamination of thin sections prepared for studies by Latter (1998) and Marsaglia et al. (1999) showed that most samples in the interval from 400 to 600 m at Site 978 consist of loose foram sand. However, two samples at 490.91 m (978A-32R-4, 11–13 cm) and at 509.99 m (978A-34R-3, 149–150 cm) are cemented by carbonate. Also, as noted by Comas et al. (1996), there is a bioclastic layer at 490.92 m (978A-32R-4, 11–13 cm) and an unusual occurrence of fish debris at 511.97 m (34R-4, 147–150 cm). The possible lag deposits (bioclasts and forams), the carbonate cementation (hardground?), and phosphatic debris suggest that one or more subtle submarine hiatuses are present from 490 to 512 mbsf.

The two thinner-than-expected intervals in Subunit IC (lower Pliocene) span periods when global sea level was high (Fig. 9). There are two possible explanations for this association. First, according to Haq (1991), there is greater dissolution of carbonate sediment during sea-level highstands. Thus these thinner-than-expected intervals could be the product of lower carbonate accumulation rates. However, there is no apparent decrease in microfossil preservation across these intervals. Furthermore, if these trends were related to eustatic highstands, then one might also expect to see the same effects in sections from other parts of the Mediterranean. As discussed above, the sections in southern Sicily and Crete are relatively complete according to published descriptions; analysis of electric log patterns from the Atlantic Ocean also shows these sections to be relatively complete (Sierro et al. 2000). Perhaps the missing section at Site 978 was caused by enhanced activity of bottom currents during highstands. This is plausible, given the likelihood of increased flow across the sill between the Mediterranean Sea and the Atlantic Ocean (at Gibraltar?) during periods of higher sea level. Thus regional conditions may have enhanced the eustatic signal at this site.

Mass Wasting During the Middle Pliocene (Subunit IB)

Various lines of evidence support the hypothesis that Subunit IB (homogeneous unit) at Site 978 is a product of mass wasting. First is the presence of slump deposits at the equivalent section at nearby Site 977. Second, Subunit IB has the characteristics of unifites as described by Stanley (1981, 1985). Compared with underlying and overlying units, Subunit IB is fairly structureless (Fig. 4B) and is both texturally and compositionally (Fig. 6) homogeneous. Stanley (1981, 1985) has documented unifite deposits within shallow piston cores in the Alboran Basin and attributed them to gravity flows. These flows are caused by sediment failure on slopes triggered by increased sediment loading during sea-level lowstands, seismic events, or tsunamis. Subunit IB at Site 978 is anomalously thick compared to the unifite layers described by Stanley (1981, 1985); the latter range up to 10 meters in thickness, whereas Subunit IB at Site 978 is over 60 meters thick. Localized bedding and bioturbation in this unit may reflect amalgamation of several separate unifite layers.

On a broader scale, evidence for mass wasting at approximately 3 Ma occurs at many other Mediterranean drill sites and onshore sections (see Table 2 and descriptions of onshore sections above). The mass-wasting features include: (1) slump scars (hiatuses), (2) semi-coherent to structurally deformed slump deposits exhibiting inclined bedding, broken bedding, small-scale faulting, failure plane(s), tightly folded sediment, coherent blocks in a muddy matrix, and intraclastic gravel or conglomerate, and (3) structureless (fluidized or homogenized) mud or unifites. Slump scars on the slope (e.g., contact between Trubi and Narbone formations) may correspond with slump deposits (e.g., Subunit IB at Site 977) and/or unifites (e.g., Subunit IC at Site 978) at the base of slope and on the basin floor.

The ubiquitous evidence for mass wasting at approximately 3 Ma across the Mediterranean leads us to believe that these features are mainly eustatic rather than tectonic in origin. These features are likely the product of slope degradational processes associated with a series of rapid sea-level fluctu-

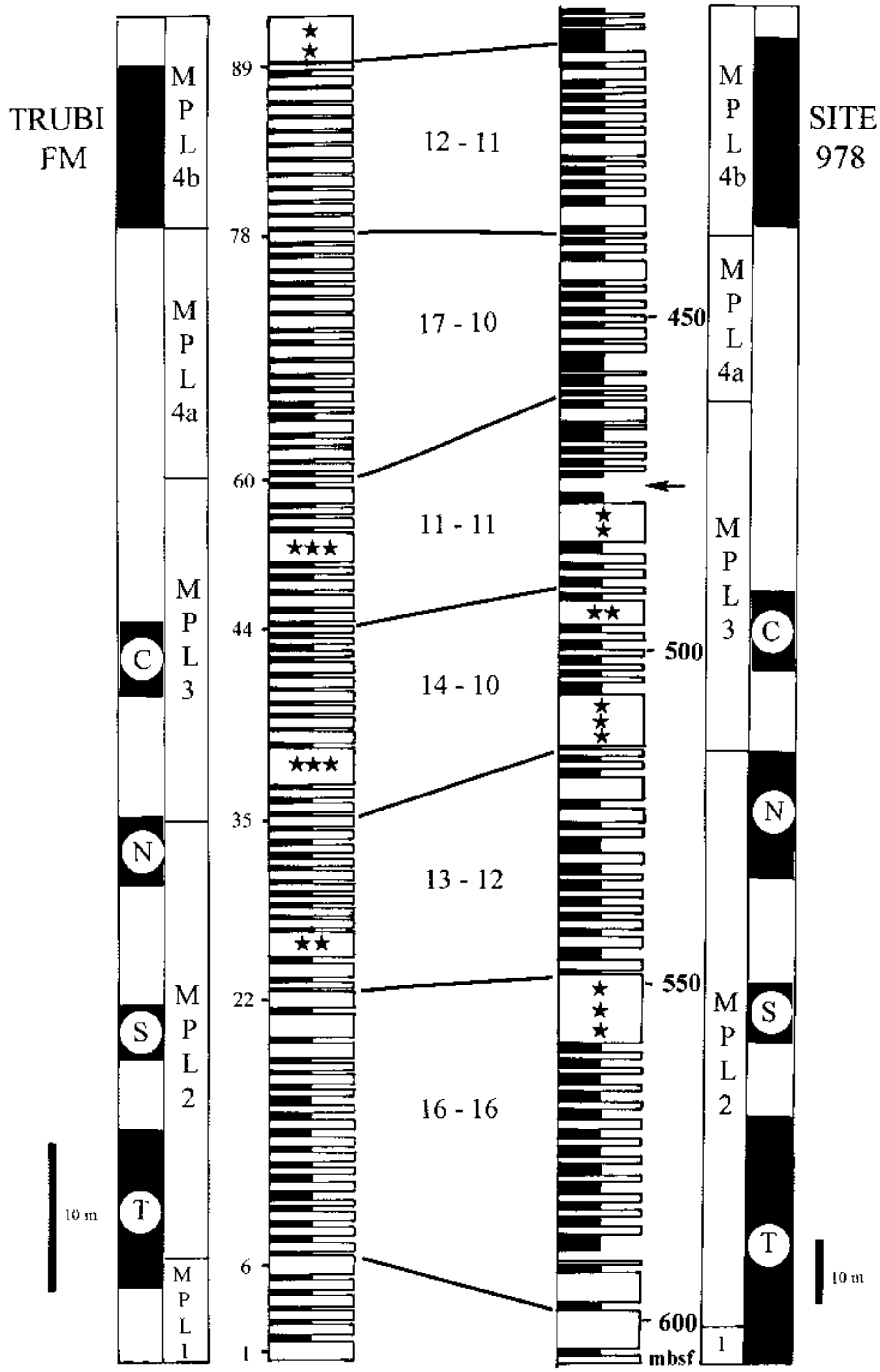

FIG. 8.—Correlation scheme for the Sicilian Trubi Formation (outcrop) and Subunit IC at Site 978 (core). To simplify correlation, beige and white marls in the Trubi cycles are grouped and assumed to correspond to simple light–dark couplets at Site 978. Lines connect biostratigraphic or magnetostratigraphic datums (top MPL1 = 5.0 Ma; top S = 4.80 Ma; top MPL2 = 4.48 Ma; top C = 4.30 Ma; top MPL3 = 3.94 Ma; top MPL4a = 3.57 Ma; = 3.33 Ma). Key beds in the Trubi column are numbered (e.g., 89, 78, 60, etc.) according to the scheme outlined in Hilgren and Langereis (1989). Numbers of cycles in opposing columns are given between datums, Trubi on the left and Site 978 on the right. Note the different vertical scales for each column; the total thickness of Trubi Formation illustrated is approximately 90 m, whereas the equivalent Site 978 section is approximately 200 m. Also note that during this correlation process we found several errors in the preliminary results volume: the core photographs for 33R and 34R (p. 864–865, Comas et al. 1996) should be switched; and the smear slide data (Section 4, p. 949–987, Comas et al. 1996) are missing depths, so that the component headings should be shifted over one column. See text for discussion.

ations in the late Pliocene (Haq et al. 1988) that began with a drop in sea level of approximately 100 m (from +70 m to −25 m) at 3.8 Ma, followed by a rise of roughly equal magnitude at 3.4 Ma and then by a drop of approximately 135 m (from +60 m to −75 m) at 3.0 Ma (Fig. 9). It is noteworthy that at 3 Ma there is evidence for a period of extreme warmth (Haywood et al. 2002) that is linked to a controversial deglaciation of the East Antarctic ice sheet as summarized in Barrett et al. (1992) and Burckle (1995). Others have noted warming in the Mediterranean region just prior to 3 Ma, as indicated by carbonate-platform development (Capozzi and Picotti 2003) and vegetation (Suc 1984). If, at 3 Ma, global warming translated to increased bottom water temperatures, then the 135 m drop in sea level may have increased methane hydrate instability and resulted in slumping episodes like the Quaternary scenarios described and summarized by Kennett et al. (2003). The latter include the Amazon fan (Maslin et al. 1998), whose mass-transport deposits, even where significantly deformed and brecciated, exhibit similar compaction trends to nontransported sediment with the exception of an overcompacted interval interpreted as a possible shear zone (Audet 1998). Workers in both the Gulf of Mexico and

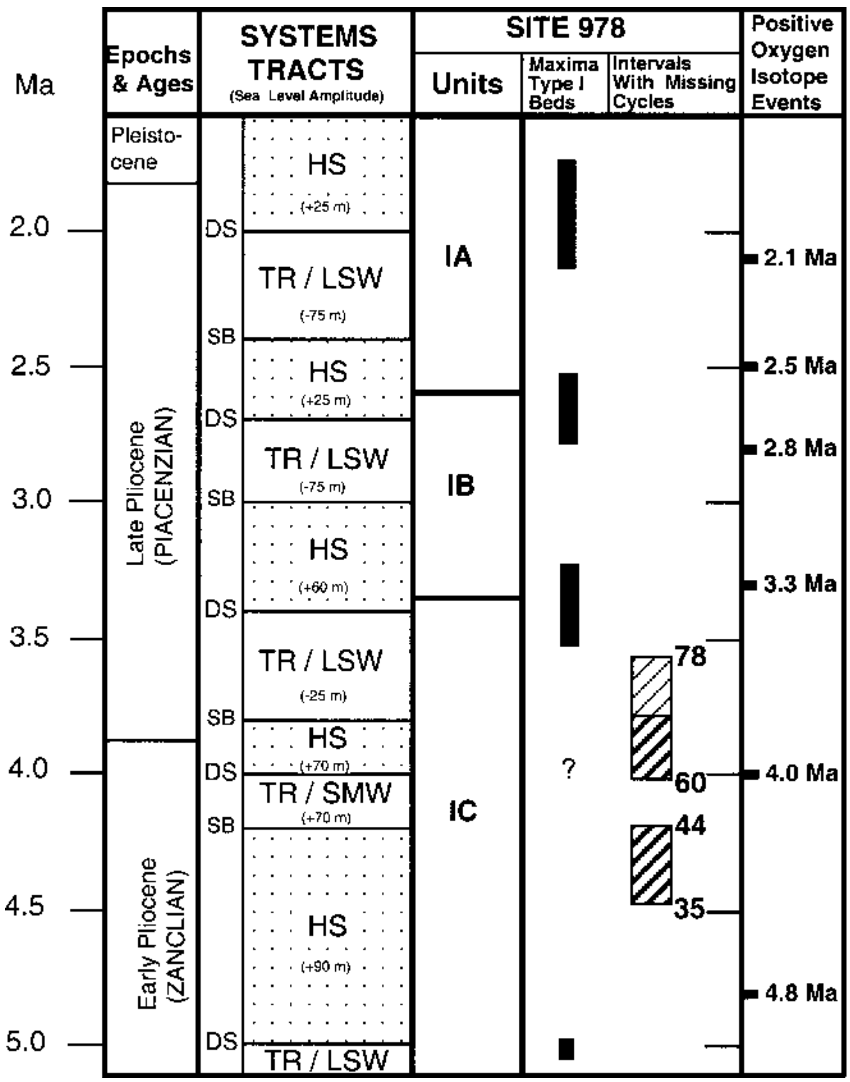

FIG. 9.—Correlation chart of significant trends in Site 978 Pliocene cycles with systems tracts and amplitudes (in parentheses) from the global sea-level curves of Haq et al. (1987, 1988). TR = transgressive systems tract; LSW = lowstand systems tract; SMW = shelf-margin-wedge systems tract; HS = highstand systems tract; DS = downlap surface. Also shown are the positive oxygen isotope events of Abreu and Anderson (1998).

the Mediterranean recognize a major sequence boundary at 3 Ma (Catalano et al. 1998), but a cursory review of DSDP and ODP cores at Atlantic Ocean margin sites that recovered middle Pliocene (3 Ma) sediments reveals no widespread evidence for mass wasting. The apparent localization of these mass-wasting effects in the Mediterranean Sea could be a function of its relatively narrow continental shelves, steep continental slopes, sediment composition and depositional history, and high regional seismicity in combination with rapid sea-level change. On the basis of our observations at Site 978 and the sedimentary record across the Mediterranean, we favor a eustatic control on middle Pliocene deposition.

Turbidite (Type 1 Cycle) Frequency in Subunit IA at Site 978

The upsection change in dominance from Type 2 to Type 1 beds and concomitant increase in sediment accumulation rate at Site 978 can be linked to the onset of Northern Hemisphere glaciation. As discussed above, a similar change in sedimentation rate is apparent in the Trubi–Narbone section. This glacial (cooling) event at approximately 3 Ma is marked by a eustatic sea-level drop of approximately 135 m, according to the curve of Haq et al. (1988), and distinct faunal changes in the Mediterranean region including benthic and planktic foraminifer and tropical Mollusca extinctions (Sprovieri and Barone 1982; Raffi and Marasti 1982; Thunell and Williams 1983; Rio et al. 1984; Zachariasse et al. 1989).

Broader-scale eustatic control on the occurrence of the Type 1 (turbidite) beds is also indicated by changes in their downhole frequency distribution (Fig. 5). Maxima in numbers of Type 1 beds occur at 240–260, 320–340, 400–420 and 580–600 mbsf. The age constraints on these maxima, as derived from Comas et al. (1996; their table 3, p. 378), are 1.75–2.13 Ma, 2.52–2.78 Ma, 3.22–3.52 Ma, and 4.98–5.10 Ma, respectively. The ages bracket those of downlap surfaces (DS) associated with all but one of the major third order cycles in the Pliocene (2.0, 2.7, 3.4, and 5.0 Ma; Fig. 9) as defined in Haq et al. (1988). Notably, there is no corresponding peak in turbidite maxima corresponding to the age of the downlap surface at 4.0 Ma. The interval where this peak would be expected, however, corresponds to an interval of missing cycles and possible missing section (see discussion of Subunit IC above).

Frequency maxima could be produced by an increase in the rate of turbidite deposition, a decrease in the rate of background sediment deposition, or a cumulative decrease in the thickness of turbidite beds. There are no apparent trends in bed thickness that correspond with these maxima. Therefore, we believe that the maxima shown in Figure 5 are actually a product of generally subtle decreases in sediment accumulation rates and represent condensed sections associated with downlap surfaces. We use the term *subtle,* because these are not of sufficient magnitude to show up on the sediment-accumulation curve (Fig. 3) within the age constraints available to us. These turbidite maxima might be the very thin, distal fringes of ''lowstand wedge'' deposits underlying downlap surfaces. Such a relationship cannot be substantitated in that the only pre–site survey multichannel seismic profile across the site (Watts et al. 1988; Comas et al. 1996) is parallel to depositional strike. This line shows subparallel reflectors within the Pliocene interval but no evidence of onlap.

Condensed sections associated with downlap surfaces can be characterized by highly radioactive (high gamma), organic-rich (high total organic carbon [TOC]) shales (Posamentier and James 1993). Maxima in TOC from shipboard geochemical analyses at Site 978 do not correspond with the spikes in turbidite frequency documented in this study. Unfortunately, the hole was not logged, so there are no data to assess relative gamma intensity of the intervals.

We can only speculate as to the cause of the steady increase in turbidite frequency maxima after 3.0 Ma (Fig. 5; 10 to 18 to 27 events at 400–420 to 320–340 to 240–260 mbsf, respectively). This trend might reflect larger-scale eustatic effects or general progradation of the fan system in response to tectonic uplift of the Betics during crustal shortening (see discussion of tectonic history in Comas et al. 1999).

Our findings appear to support those of Kennett et al. (2003), who link higher rates of turbidite deposition in the Quaternary to climate transitions during glacial termination rather than to glacial maxima or minima. However, they tie this phenomenon to clathrate disassociation, whereas we have no evidence for this during deposition of turbidite maxima in the Alboran Basin.

COMPARISON WITH REGIONAL SEQUENCE STRATIGRAPHIC STUDIES

Our model for the Pliocene units described at Site 978 is consistent with the regional seismic-sequence model of Alonso and Maldonado (1992). They subdivided the post-Miocene sediments in the Alboran Sea into three sequences: (1) basal deposits characterized by a lack of internal reflectors, (2) a middle sequence that includes mass-movement deposits, and (3) an upper sequence of stacked progradational clinoforms, submarine canyons, channel-levee complexes, and turbidite depositional lobes. They proposed that the lowermost sequence consists of fine-grained sediments deposited during the early Pliocene global eustatic highstand, and that the middle sequence records global sea-level change in the late Pliocene. This transition marks the shift to the sandy lowstand deposits that constitute the uppermost sequence that extends through the Quaternary.

This model differs from that of Stanley (1985), who suggested that the main control on Pliocene sedimentation in the Mediterranean was tectonic activity. He stated that there is ample evidence for structural deformation of Pliocene basin-margin sediments throughout the Mediterranean. Possible examples are provided by Channell et al. (1994) and Sartori (1990), who attributed hiatuses (and deposits associated with mass wasting) in the middle Pliocene (MPL4 and MPL5) sections in the Northern Apennines of Italy to tectonic activity. Although this period of middle Pliocene tectonism is well documented in the central Mediterranean, there is no other information that supports increased tectonic activity during this period throughout the entire Mediterranean.

However, there is evidence that eustasy can play *the* major role in the supply of terrigenous sediment to deep marine settings in tectonically active areas (e.g., Ito and Katsura 1992; Pickering et al. 1999). Several workers have tried to identify the record of sea-level change in the Pliocene of the central Mediterranean from field and subsurface studies. Capozzi and Picotti (2003) related carbonate-platform development, sapropel development, and to a lesser extent, turbidite deposition in the Northern Appenine foredeep to climate and eustatic changes. Catalano et al. (1998) defined depositional sequences in pelagic Pliocene facies based on the identification of unconformities (sequence boundaries) and maximum flooding events, which they defined by thicker carbonate and shale/sapropel intervals (condensed intervals). According to Catalano et al. (1998), thicker carbonate units (combination of obliquity and precession cycles) dated at 2.0, 2.2, 2.8, 3.65, and 4.4 Ma represent condensed sections deposited during global warm periods, whereas thicker shale and sapropel units dated at 1.75, 2.4, 2.95, and 3.82 Ma represent condensed sections deposited during cooler periods. Some of these overlap the age distribution of turbidite maxima at Site 978 (1.75–2.13, 2.52–2.78, 3.22–3.52, and 4.98–5.10 Ma), but there is no direct correspondence. They also overlap with the ages of condensed sections recognized in the Gulf of Mexico. For example, Crews et al. (2000) reported carbonate-rich condensed sections (CRCS) and shale-rich condensed sections (SRCS) with approximate ages of 1.95 (CRCS), 2.3–2.85 (CRCS), and 3.6 (SRCS). As discussed in Catalano et al. (1998) chronostratigraphic differences among these data sets could be linked to their use of different time scales and biochronozones. Because Catalano et al. (1980) did not report the detailed basis for their determinations and admittedly inferred the presence of sequence boundaries and condensed sections in outcrop, we can only report this discrepancy with our observations at Site 978 and leave its resolution to future work.

COMPARISON WITH OXYGEN ISOTOPE CURVE

Many have questioned the reliability of the Haq et al. (1987) curve and sequence stratigraphy outlined in Figure 9. Oxygen isotope data have been proposed as more realistic proxies for sea-level change as the isotopic signature relates to glacial ice volume (e.g., Shackleton et al. 1995). However, interpretations of the Pliocene oxygen isotope record are also controversial and open to interpretation (Burckle 1995). For example, Shackleton (1995) and Shackleton et al. (1995) question whether some light isotopic excursions in the Pliocene record reflect Antarctic ice reduction or warming of the deep oceans. Abreu and Anderson (1998) constructed a composite isotopic record with the Pliocene section based on foraminifera data from Site 502 in the Caribbean (Oppo et al. 1995) and Site 704 in the southern Atlantic Ocean (Hodell and Venz 1992). They identify several positive isotope events in the Pliocene, mostly high-amplitude isotopic events, which according to Abreu and Anderson (1998) would have corresponded to sea-level variations up to 90 m owing to ice-sheet growth. These events are plotted in Figure 9. It is interesting that the pattern of maxima in Type 1 beds corresponds more to the system-tract boundaries both in number of events and relative timing than to the oxygen isotopic events. Thus our data appear to validate the Pliocene third-order cycles of Haq et al. (1987).

SUMMARY AND CONCLUSIONS

Prior to Leg 161, rhythmically bedded Pliocene units in the Mediterranean had primarily been described from outcropping sections in Sicily, mainland Italy, and Crete. Site 978 in the Alboran Sea provides an excellent Pliocene sedimentary record that can be confidently correlated with other sections in the Mediterranean region. The three Pliocene subunits at Site 978 are consistent with the proposed tripartite subdivision of the Pliocene into upper, middle, and lower series and they likely reflect climate and sea-level change within the Mediterranean region. There is good correlation between cycles in the rhythmically bedded lower Pliocene (Subunit IC) of Site 978 and classic Pliocene outcrops in Sicily and mainland Italy. These cycles most likely reflect changes in continental runoff associated with astronomical cycles in precession and obliquity. Missing cycles correlate with eustatic highstands. We attribute the middle Pliocene shift in depositional style (Subunit IB) to rapid sea-level change, and subsequent large-scale submarine mass wasting. There is evidence from correlative middle Pliocene units in other ODP cores and outcrop sections that this was a Mediterranean-wide event. This period also marks the onset of Northern Hemisphere glaciation and a series of late Pliocene glacio-eustatic changes in sea level that can be tied to periodic influx of clastic sediment at Site 978 and to a post–3 Ma increase in sediment accumulation rates throughout the Mediterranean region. Further eustatic control on sedimentation can be seen in the apparent correlation between maxima in turbidite frequency and the age of downlap surfaces and associated condensed sections.

The identification of downlap surfaces on seismic lines and maximum flooding surfaces in sedimentary successions is one of the first steps in a sequence-stratigraphic study (Posamentier and James 1993). These surfaces are key chronostratigraphic markers used in sequence stratigraphic correlations (Loutit et al. 1988; Galloway 1989; Van Wagoner et al. 1990; Liu et al. 1998). Although downlap surfaces can be easily deciphered on low-resolution seismic lines, their definition in the rock record is more ambiguous and based largely on lithology (e.g., Carter et al. 1998; Crews et al. 2000). Herein we have outlined an additional criterion for discerning condensed sections associated with downlap surfaces, heretofore unrecognized in the rock record, that of relative maxima in turbidite frequency. We propose that this signal results from decreased rates of background sedimentation over a period of uniform turbidite deposition, or alternatively represents the ''condensed'' fringe of a lowstand-wedge deposit. Recognition of the apparent link between maxima in turbidite frequency and downlap surfaces (condensed sections) may be an important tool in discerning eustatic signals in distal settings of ancient deep-marine sequences. It may be that this sort of signal is characteristic of more distal settings already dominated by mudstones with higher carbonate content. Unfortunately the literature provides little basis for comparison, in that prior statistical studies of turbidites have focused on trends in bed thickness and grain size (e.g., Chen and Hiscott 1999; Ishihara et al. 1997). Nevertheless, this relationship is a hypothesis that bears further testing.

This study is significant because eustatic effects have been reported mainly from extensional basins (Miall 1996), where they are most readily recognized, whereas the Alboran (Betics) region was characterized by compressional tectonics during the Pliocene (Comas et al. 1999). Deciphering the eustatic signals was made possible through a combination of high-resolution biostratigraphy, magnetostratigraphy, and cyclostratigraphy at this site.

ACKNOWLEDGMENTS

We thank Michael Arthur, David Piper, and William Sager for their helpful comments on early versions of this paper, as well as associate editor Michael Underwood, and reviewers Craig Fulthorpe and Robert Carter, who provided excellent suggestions that were incorporated into this final version. This research used data provided by the Ocean Drilling Program (ODP), which is sponsored by the U.S. National Science Foundation and participating countries under management of Joint Oceanographic Institutions, Inc. This paper would not have been possible without the hard work and excellent data sets generated by our fellow Leg 161 Shipboard Scientists: C. Aubourg, P.E. Belanger, S.M. Bernasconi, M.C. Comas, E.P. de Kaenel, F.D. de Larouzière, C. Doglioni, H. Doose, M. Hobart, S.M. Iaccarino, P. Ippach, A. Murat, G.M. O'Sullivan, J.P. Platt, W.G. Siesser, J.I. Soto, K. Tandon, M. Torii, and J.S. Tribble.

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Received 9 May 2002; accepted 24 September 2003.