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1. TECTONIC INTRODUCTION[1]

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INTRODUCTION

Ocean Drilling Program (ODP) Leg 160 is the first in a two-leg program to investigate the tectonic and paleoceanographic history of the Mediterranean Sea (Fig. 1). It represents the fourth leg of scientific drilling in the Mediterranean, after Deep Sea Drilling Project (DSDP) Legs 13 (Ryan, Hsü, et al., 1973) and 42A (Hsü, Montadert, et al., 1978) and ODP Leg 107 (Kastens, Mascle, et al., 1990). One focus of this leg is on accretionary and collisional processes associated with the convergent boundary between the African and Eurasian plates. The other focus is on the origin and paleoceanographic significance of sapropels, organic-rich layers that are intercalated in the Pliocene-Pleistocene sediments of the Mediterranean basin.

RATIONALE FOR TECTONIC DRILLING

The African/Eurasian plate boundary in the Eastern Mediterranean region reflects tectonic settings ranging from effectively steady-state subduction to incipient collision and more advanced collision in different areas (Fig. 1). The Eastern Mediterranean is, thus, an ideal area to investigate the transition from subduction to collision, processes that may be recorded on land in orogenic belts, but which are presently poorly understood (e.g., McKenzie, 1972, 1978; Le Pichon, 1982; Robertson and Grasso, 1995). Such early collisional settings have not been studied in any detail by the ODP to date, despite their obvious importance to earth science. The entire Mediterranean basin is surrounded by deformed lithologic units that originated within the Tethyan ocean (Fig. 2). Unlike inactive orogenic belts (e.g., Iapetus), there is an opportunity in the Mediterranean region to link observations from deep-sea drilling with those from earlier collisional events that are recorded on land, in mountain chains, including the Dinarides, Hellenides, and Taurides (e.g., Robertson, 1994).

The present "shallow phase" of Mediterranean drilling was aimed at focusing on the Mediterranean Ridge, an example of a mud-dominated accretionary complex (Fig. 1). However, only two transects, the Ionian transect (Sites 971–973) in the west and the Eratosthenes Seamount–Cyprus transect could be drilled during Leg 160 in view of political and logistical constraints. In practice, the most easterly of the original transects proposed, Eratosthenes Seamount–Cyprus, featured strongly in Leg 160. The sites selected for drilling of sapropels were selected independently of any tectonic interest. However, one of these sites (Site 964), on the Pisano Plateau, is located close to the deformation front of the Calabrian accretionary wedge and yielded additional information of tectonic interest. Also, Site 963 is located on the unstable foreland of the African plate, south of Sicily, in an area affected by thrust loading.

Leg 160 included the study of geological responses to collisional processes in three contrasting tectonic settings. The first of these settings, in the east, is the locus of collision of an inferred continental fragment, the Eratosthenes Seamount, with the Eurasian continental margin to the north, represented by the Cyprus active margin. A north-south transect of four holes was made from the Eratosthenes Seamount to the Cyprus slope. The second objective concerned processes related to the formation of mud domes, resulting from mud diapirism or mud volcanism on the Mediterranean Ridge, within the Olimpi area, south of Crete. Three holes were drilled on the Milano mud dome (Site 970), one on the crest and two on the flanks. A further four holes were drilled on the Napoli mud dome (Site 971); one on the flanks, one near the base of the dome, and two on the crestal area. In addition, the eastern flank of the Milano Dome in the Olimpi Field was drilled, because the seismic character of the flank facies was clearly imaged there. Some additional information on the regional tectono-sedimentary setting of the Pliocene-Pleistocene of the Mediterranean Ridge was obtained by drilling nearby for sapropels at Site 969. The third objective, in the west, was the Ionian deformation front, where a transect of three holes was drilled, one on the incoming abyssal plain and two on the lower and mid-slopes of the accretionary prism. The aim was to sample the incoming sediments and then to compare them with the accreted material, in which fluids may have been affected by salt tectonics associated with the Messinian desiccation event.

REGIONAL GEOPHYSICAL SETTING

Subduction of the African plate with respect to Eurasia is nearly orthogonal in a northeast direction in the Ionian Basin, at a low angle toward the northwest in the western Levantine Basin (with prevailing left-lateral strike slip), and again nearly orthogonal in the easternmost Levantine Basin along an active margin near Cyprus (McKenzie, 1972, 1978; Le Pichon et al., 1982a, 1982b). Prior to Leg 160, the plate tectonic framework of the Calabrian and Mediterranean accretionary prisms was better understood than the more easterly segment that includes the Eratosthenes Seamount and Cyprus.

The crustal thickness of the incoming plate indicates that oceanic crust, probably of Mesozoic age, persists to the southwest of the Mediterranean Ridge beneath the Ionian Basin and the Sirte Abyssal Plain, whereas crustal thickening occurs beneath the Mediterranean Ridge and the Hellenic Trench System (Rabinowitz and Ryan, 1970; Woodside and Bowin, 1970; Finetti, 1976; Giese et al., 1982; Le Pichon et al., 1982b; Makris et al., 1983; Makris and Stobbe, 1984; Underhill, 1989; Kastens, 1991; Kastens et al., 1992; de Voogd et al., 1992; Truffert et al., 1993). Farther east, the crust is thicker between the Cyrenaica Promontory and the island of Crete, in an area where no abyssal plain is present south of the deformation front and where continental collision is taking place. The Mediterranean Ridge is bounded to the north by the Hellenic Trench System and the island of Crete, which can be interpreted as a forearc high. Farther north is the South Aegean Volcanic Arc, including Santorini volcano (Fytikas et al., 1984), and then the Aegean backarc basin system (Horvath and Berckhemer, 1982; Kissel and Laj, 1988).

Seismological evidence also defines the Africa/Eurasia plate boundary in the Eastern Mediterranean. A broad zone of intermedi-

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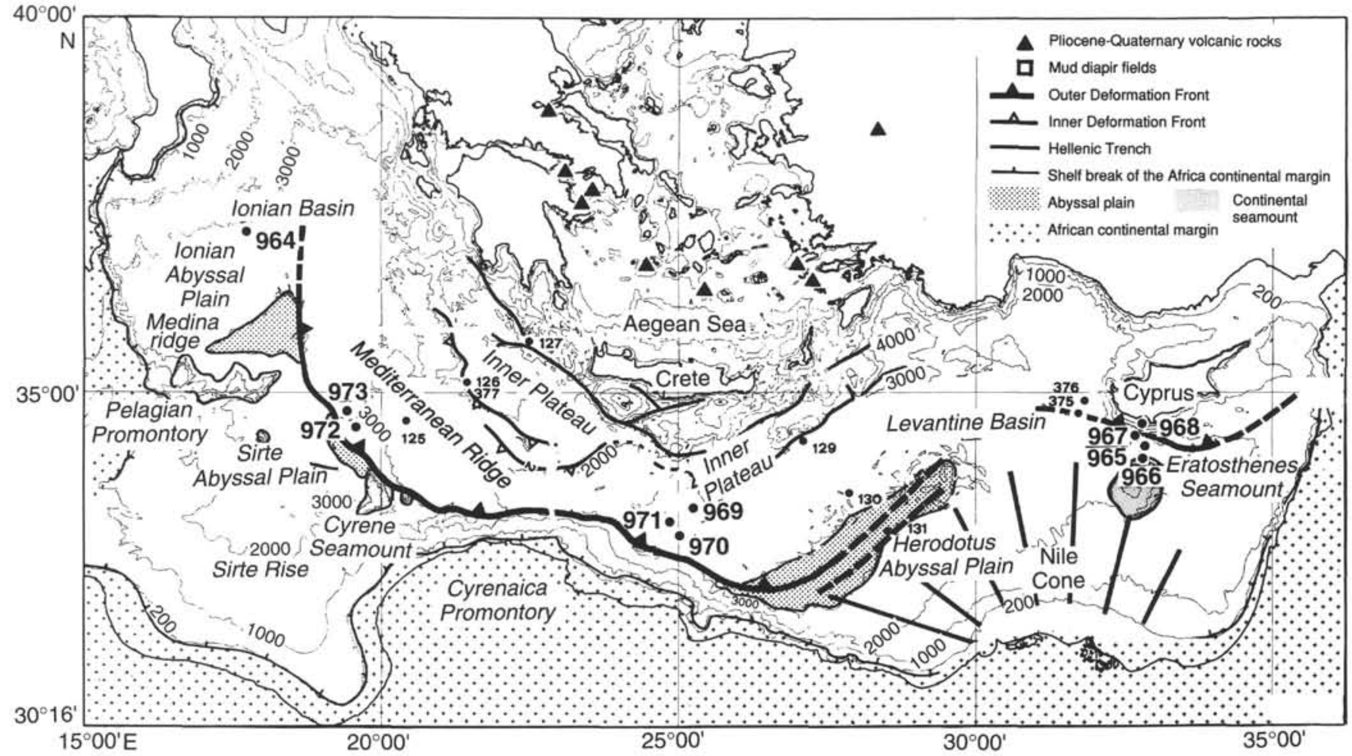

Figure 1. Outline map of the Mediterranean showing the main tectonic features and location of the sites drilling during Leg 160. DSDP sites are also shown.

ate-depth earthquakes (up to 200 km deep) delineates a generally northward-dipping subduction slab, especially along the Hellenic Trench System (Wortel et al., 1990). The timing of the onset of subduction along the active margin system in the Eastern Mediterranean is still debated. The geophysical evidence (e.g., from seismic tomography) is ambiguous. However, field geological evidence from Crete (Meulenkamp et al., 1988, 1994) and Cyprus (Eaton and Robertson, 1993) suggests an early Miocene age for the start of the cycle of northward subduction.

In the easternmost Mediterranean, between the island of Rhodes and the Levant coast, the present-day and past history of subduction are less well defined than farther west (Fig. 3). A combination of seismic, magnetic, gravity, and bathymetric evidence indicates that a northward- or northeastward-dipping subduction zone is present south of Cyprus. To the west of Cyprus, northeastward subduction is indicated by earthquake data (Rotstein and Kafka, 1982; Kempler and Ben-Avraham, 1987) and is also revealed by crustal shortening along the Florence Rise and in the Antalya Basin, where folding, thrusting, and northeastward tilting and deepening are observed (Woodside, 1976, 1977). The seismicity of the active margin to the south of Cyprus is less than that associated with the Aegean arc, but it indicates subduction oriented north-south and north northwest–east southeast (Rotstein and Kafka, 1982; Ben-Avraham and Nur, 1986; Woodside, 1991). Eastward, the zone of convergence inferred to lie between Cyprus and the Eratosthenes Seamount provided the basis for a focus of drilling during Leg 160. To the east of Cyprus, dominantly strike-slip motion with some component of shortening is inferred by most workers (Kempler and Ben-Avraham, 1987; Kempler, 1993), although the exact location of the plate boundary and how it connects with the East Anatolian Fault and the Bitlis Suture Zone in eastern Turkey remain unclear. The seafloor shallows toward the Levantine coast, where it is cut by north-east-trending escarpments. Another prominent tectonic lineament, the Kyrenia Range, runs east-west, through the northern part of Cyprus. Eastward, this connects with the Misis Mountains in southern Turkey (Kelling et al., 1987). On either side of this lineament, the Cilicia, Adana, and Iskenderun basins contain thick Pliocene-Pleistocene sediments and are inferred to be extensional and strike-slip- (i.e., trans-tensional) controlled basins (Sengör et al., 1985; Kempler and Garfunkel, 1991).

In the following section, we discuss the tectonic setting of each of the three areas of the Africa-Eurasia convergent zone that were targeted for drilling during Leg 160.

ERATOSTHENES–CYPRUS MARGIN TRANSECT: INCIPIENT CONTINENTAL COLLISION

The main objective was to study processes of incipient collision of continental crust along the easternmost segment of the active margin separating Africa and Eurasia in the Mediterranean Sea. Processes of continental collision are complex and remain poorly understood. For example, the converging continental margins are not linear and, as a result, opposing margins may vary in shape and crustal thickness (Dewey, 1980). Passive margins may include continental fragments that were rifted from a neighboring continent and that remained adjacent to a subsiding passive margin (e.g., Rockall Bank of the eastern North Atlantic Ocean). The same could be true for large igneous features (e.g., oceanic plateaus). Such elements would remain passive until much later, when they became involved in the initial stages of continental collision. This situation is applicable to the present North African continental margin, which originated by rifting of Gondwana in the early Mesozoic (or even earlier) and has remained as a passive margin until Pliocene-Pleistocene time in the Eastern Mediterranean. Collisions occurred relatively early in the case of large salients such as the Arabian Peninsula in the east and the Cyrenaica Peninsula in the central Mediterranean. Indeed, the Arabian segment of the Gondwana margin in the east represents a major promontory that finally collided with Eurasia along the Bitlis Suture Zone by the late Miocene, and which has influenced the geological history of the easternmost Mediterranean region (Mart, 1994).

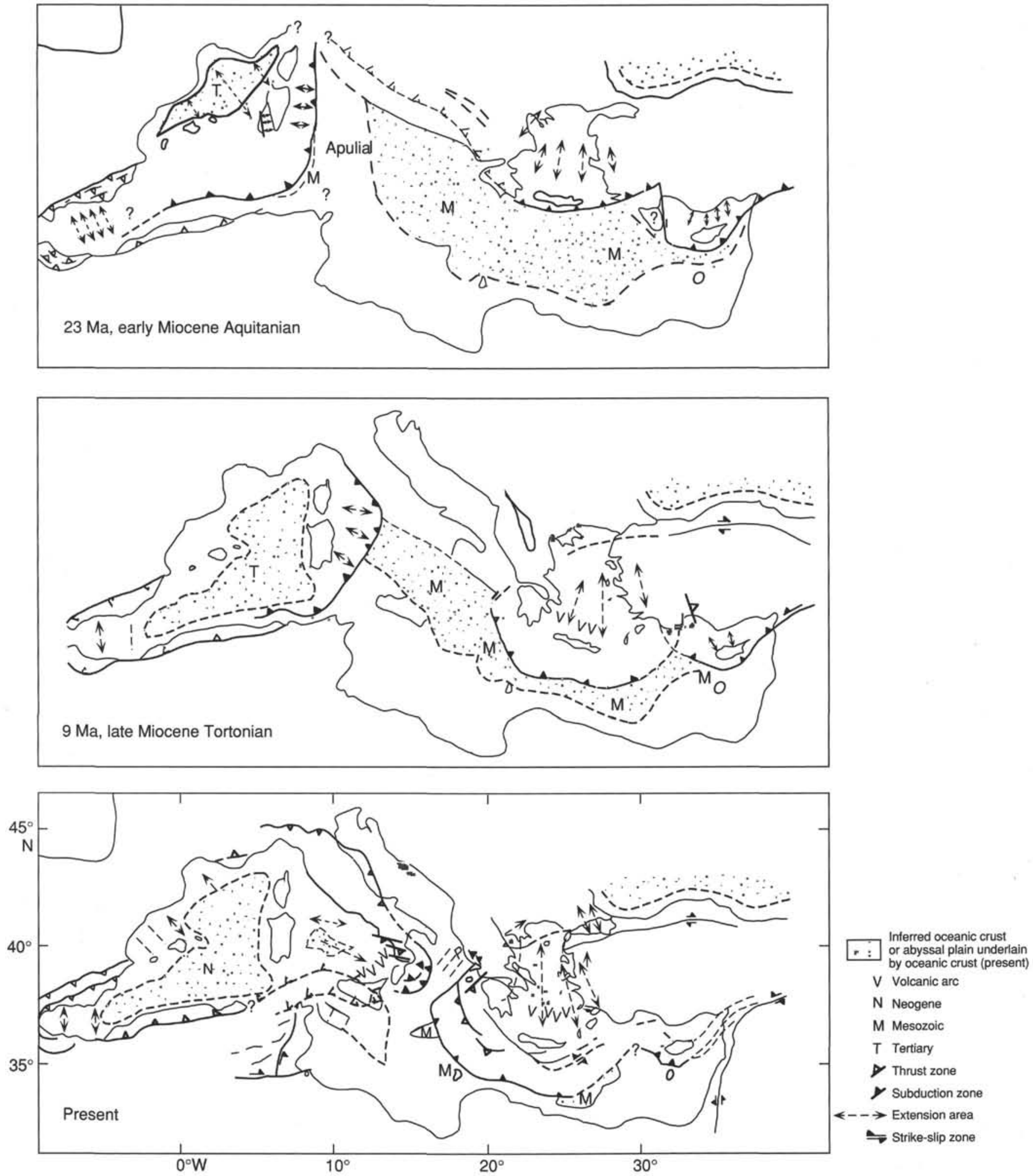

Figure 2. Schematic plate tectonic reconstruction of the Mediterranean region from Miocene to Holocene time. From Robertson and Grasso (1995). Leg 160 investigated the tectonic processes along the plate boundary that separates the African and Eurasian plates.

The Eratosthenes Seamount is the most prominent bathymetric feature between the Nile Cone and Cyprus, rising from a depth of approximately 2000 to <700 m on the crestal plateau area (Fig. 4). Seismic refraction data and gravity modeling (Woodside and Bowin, 1970) suggest that the crust thins by about 2–3 km along an imaginary north-northeast to south-southwest line through Eratosthenes Seamount from a thickness of 24–26 km in the Levantine Basin in the east (Woodside, 1977, 1991; Makris and Stobbe, 1984). The sedimentary succession could be as much as 15 km thick. The seismic velocity of the underlying crust is about 6.7 km/s, which is high for continental crust, but would be reasonable for oceanic crust (Layer 3) if it were not so thick (10 km), compared to more typical values (5

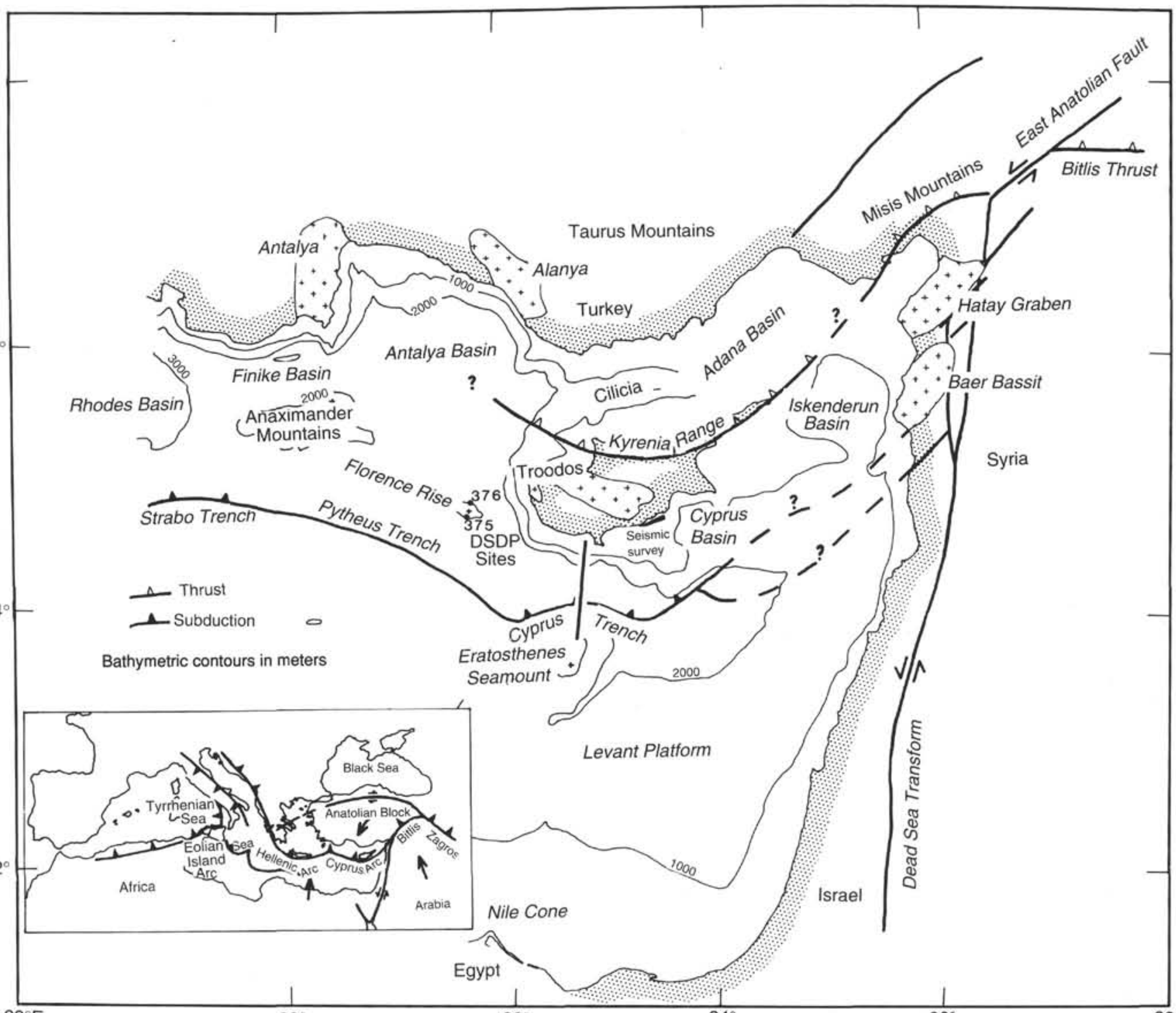

Figure 3. Schematic tectonic map showing the location of the Eratosthenes Seamount and the island of Cyprus in relation to the tectonic framework of the easternmost Mediterranean region. From Robertson et al. (1991). The line in the center shows the Leg 160 drilling transect.

km). A large magnetic anomaly is associated with the Eratosthenes structure, which has been taken to suggest the presence of a broad, deep-seated source (Ben-Avraham et al., 1976). The magnetic anomaly ranges in amplitude from −150 nT to the northwest to 450 nT to the southeast and is elongated in the same direction as the seamount (Fig. 5A). The length of the anomaly in the north-northeast–south-southwest direction is almost 300 km between enclosing 0-nT contour lines and more than 100 km in the northwest-southeast direction between the maximum and minimum values. Several superimposed small local anomalies distort the overall anomaly in the northwest. Ben-Avraham et al. (1976) interpreted the source as basic and ultrabasic (possibly ophiolitic) material. Woodside (1977) suggested that the source might be basic igneous intrusions.

In strong contrast to the magnetic anomaly pattern, the gravity anomaly is almost entirely topographic (Woodside, 1977; Fig. 5B). There is a free-air anomaly of about 75 mGal that diminishes to a Bouguer anomaly of less than 20 mGal after correction for a topographic mass with an assumed mass of 2.67 g/cm^3. This suggests that the seamount is supported by the crust alone or by forces within the upper lithosphere, as there appears to be no lower crustal root or flexural response to the topographic load. A regional westward increase in gravity by about 20 mGal just to the west of the seamount corresponds to the westward thinning of the crust by about 2–3 km. To the west of this line, there is a north-northeast–south-southwest-oriented long-wavelength linear positive gravity anomaly of 30–50 mGal, which indicates the presence of a major crustal discontinuity (Woodside, 1991; Krasheninnikov and Hall, 1994).

Bathymetric charts and the results of a swath bathymetric survey conducted over the southern part of the Eratosthenes Seamount during a cruise of *Academician Nikolai Strakhov* indicate that the Eratosthenes Seamount is approximately rectangular (Fig. 6). The southern and eastern margins are steep to locally very steep. The northern and northwestern flanks are markedly terraced. The summit area is broadly undulating, with an east-northeast–west-southwest trend. Early seismic data revealed that the northern and southern margins of the seamount are bordered by thick sedimentary basins (Ryan, Hsü, et al., 1973; Ross and Uchupi, 1977; Montadert et al., 1978; Woodside, 1977), here termed the northern and southern basins. Seis-

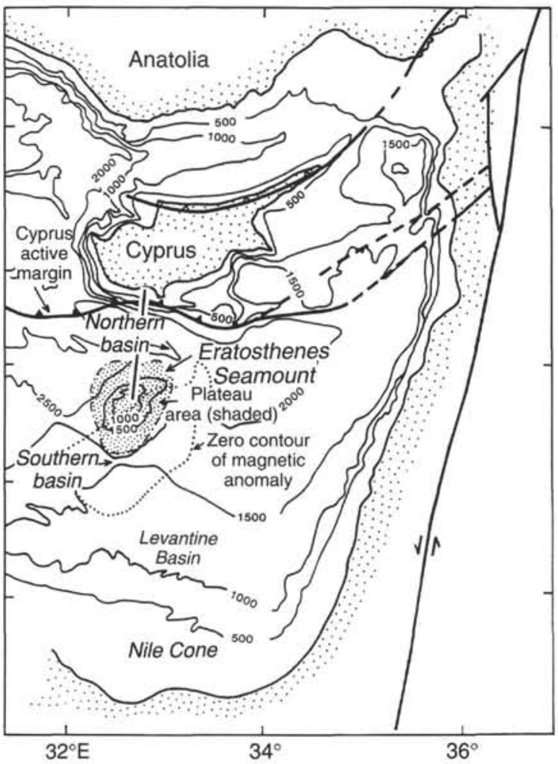

Figure 4. Tectonic setting of the Eratosthenes Seamount in relation to Cyprus. Note the location of the magnetic anomaly relative to the seamount. From Robertson et al. (1994). The line shows the Leg 160 drilling transect.

mic evidence suggests that Messinian evaporites were probably not deposited on the crest of the seamount, but they can be seen pinching out against the southern and western margins. There is no clear seismic evidence that thick evaporites were formed in the northern basin between the Eratosthenes Seamount and Cyprus (Limonov et al., 1994). Successions of nannofossil muds with sapropels and tephra were cored near the crest of the seamount at Station KC20 as part of the European Union MARFLUX program. In addition, there have been informal reports of dredging of Cretaceous limestones from the southern lip of the seamount. Recently, the samples were examined by Y. Mart (pers. comm., 1995), who was able to confirm a Miocene age.

Information obtained during the site surveys for Leg 160 during the 1993 TREDMAR-3 cruise (Limonov et al., 1994) comprised 12-channel air-gun seismics, OKEAN long-range side-scan sonar, and MAK deep-towed short-range side-scan sonar, combined with a deep-towed 5.5–5-kHz high-resolution profile; gravity and box coring were also conducted. The various lines run using these techniques are shown in Figures 7 and 8. The data obtained reveal the following main features (Robertson et al., 1994, in press):

1. The base of the inferred Pliocene-Pleistocene succession is clearly imaged for the first time and is estimated as being at a depth of <150 m throughout the crestal area.
2. The inferred Messinian reflectors pinch out against the margins of the seamount, indicating that evaporites are not present on the seamount.
3. Sediments, interpreted as Pliocene-Pleistocene turbidites, dip beneath the northern basin, suggesting that active underthrusting is taking place.
4. An approximately east-west-trending narrow ridge is near the base of the slope along the northern margin of the seamount. This ridge is separated from the lower slope of the seamount to the south by a small basin in which the sediments appear folded.
5. The northern and eastern margins of the seamount are dissected by normal faults. Bedrock is exposed along scarps (e.g., along the southern and western lips of the seamount).
6. The crestal areas of the seamount are cut by high-angle faults; some of these cut only the inferred Pliocene-Pleistocene sequence, whereas others cut the underlying unit.
7. Sediments within the southern trough appear to dip southward under the slope of the Levantine Basin, which suggests that active underthrusting of the Eratosthenes Seamount is taking place (i.e., southward as well as northward).

The location of a convergent plate boundary between Cyprus and the Eratosthenes Seamount was previously obscured by interpretations based on seismic refraction that both Cyprus and the Eratosthenes Seamount are located on continental crust (Makris et al., 1983). Another common misconception is that the plate boundary is located within Cyprus, possibly along the front of the Kyrenia Range (McKenzie, 1972; Ben-Avraham and Nur, 1986). By contrast, many other authors have recognized that the plate boundary is located between the Eratosthenes Seamount and Cyprus (Woodside, 1977; Le Pichon and Angelier, 1979; Rotstein and Kafka, 1982). Gass and Masson-Smith (1963) were far ahead of their time in suggesting that Cyprus had been uplifted as a consequence of thrusting of the African continental margin beneath Cyprus. Moores and Vine (1971) suggested that the uplift of the Troodos ophiolite was caused by diapiric protrusion of serpentinite that was formed by hydration of ultramafic rocks within the lower part of the ophiolite sequence. Studies of the sedimentary cover of the Troodos ophiolite have elucidated the timing of uplift of the ophiolite (Robertson, 1977; Poole and Robertson, 1992). More recently, it was suggested that the uplift was triggered by underthrusting of the Eratosthenes Seamount beneath Cyprus (Robertson, 1990, 1992; Kempler and Garfunkel, 1992; Robertson and Xenophontos, 1993; Robertson et al., 1994, in press; Payne and Robertson, in press). Additional suggestions are that the seamount may be displaced by thrusting toward the west and that the eastern margin may be affected by strike-slip faulting (J. Woodside, pers. comm., 1995). Indeed, the seamount could be undergoing westward "tectonic escape" and "rotation" in addition to northward underthrusting.

A contrasting picture was envisaged by Kempler (1993), who interpreted the present structure of the Eratosthenes Seamount as related to extensional rather than compressional tectonics. She inferred the seamount to be a quadrilateral-shaped graben, surrounded by deep, sediment-filled moats. Bounding steep normal faults were inferred with throws in excess of 500 m. Likewise, Y. Mart (pers. comm., 1994) also believed that the Eratosthenes Seamount is bounded by extensional faults and, further, inferred that diapiric structures located to the southeast of Eratosthenes Seamount are related to an extensional tectonic setting. Another current controversy concerns the deep structure of the Eratosthenes Seamount. One view is that this is probably continental crust, possibly with dense igneous intrusive bodies at depth (Woodside, 1977, 1991), while another is that the seamount represents thick oceanic crust, similar to an oceanic plateau, with only a relatively thin sedimentary cover (Y. Mart, pers. comm., 1995). The working assumption during Leg 160 was that the seamount is essentially a continental structure related to rifting of Gondwana in early Mesozoic time (Kempler, 1993) and that it is currently being deformed by collision of the African and Eurasian plates (Robertson et al., 1994, in press; Limonov et al., 1994).

Between Miocene and Holocene time, the Eratosthenes Seamount entered a particularly interesting phase in which the leading, northern, edge of the seamount has been in the process of collision with the Cyprus active margin, causing break-up and collapse (Limonov et al., 1994; Robertson et al., 1994, in press). The southern flank is also ap-

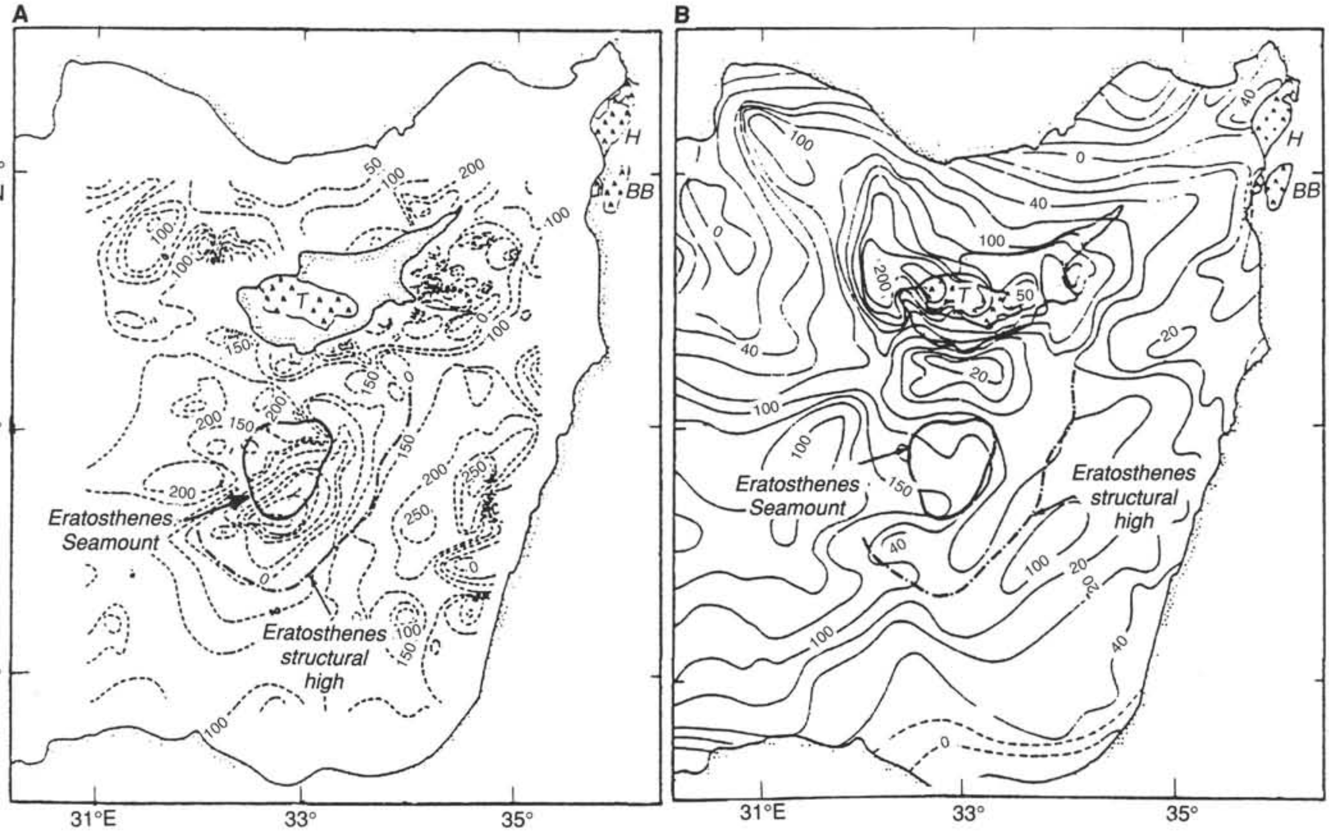

Figure 5. Magnetic (**A**) and Bouguer (**B**) anomalies (nT and mGal, respectively) in the easternmost Mediterranean, simplified after Woodside (1976) and redrawn from a simplified version in Kempler (1993). BB = Baer Bassit, H = Hatay Graben, T = Troodos ophiolite.

parently beginning to subside and be thrust beneath the sediments of the Levantine Basin, although the kinematic setting of this area remains poorly documented. The collision of Eratosthenes Seamount, or its former northward prolongation, appears to have had an important influence on the uplift of Cyprus in the late Pliocene–Pleistocene (Robertson, 1990). The stage was thus set for the documentation of collisional tectonic processes by drilling a north-south transect of four sites across the plate boundary between Cyprus and the Eratosthenes Seamount during Leg 160. Similar early collisional processes must have taken place in many mountain belts on land (e.g., Alpine-Mediterranean Tethys), but the effects have generally been obscured by later collisional deformation, including uplift, large-scale thrusting, and folding (Robertson, 1994).

MUD DOMES ON THE MEDITERRANEAN RIDGE

The major objective of drilling mud domes on the Mediterranean Ridge (Fig. 1) was to try to distinguish between alternate possible origins involving either "mud diapirism" or "mud volcanism." The tectonic setting of the Mediterranean Ridge is shown in Figure 1, and recent geophysical interpretations of the structure of the ridge are shown in Figures 9 and 10. Mud domes in this area (Fig. 11) can be considered as protrusions (i.e., upward movements) of material, which could have formed pluglike bodies that were forced upward to form raised features on the seafloor. On the other hand, mud volcanoes could result from the eruption of relatively fluid mud from a central vent, followed by outward flow to build up a series of mudflows, analogous to the formation of subaerial volcanoes (Figs. 12, 13). If the diapiric hypothesis were correct, we would expect to drill a relatively homogeneous pluglike mass of mud breccia; on the other hand, if the mud volcano-type hypothesis were correct, we would anticipate the occurrence of distinctive mudflows (i.e., debris flows), intercalated with background deep-water sediments on the flanks of the mud mounds. In reality, these two processes need not be mutually exclusive, as mud may be emplaced upward as a true diapir at depth and then ejected as more fluid mud at the seafloor, building up conical structures. Also, not all mud domes are necessarily of identical origin. Below, we summarize some of the available information on the mud domes.

Mud domelike stuctures have now been extensively identified in many areas of the Mediterranean, including both the eastern and western basins, and they are also present in the Black Sea (e.g., Limonov et al., 1994). On the central part of the Mediterranean Ridge that was drilled during Leg 160, mud domes were identified by single-channel seismic profiling and by gravity coring in a small area of the northern edge of the "upper plateau," on the flat crestal area of the ridge (Cita et al., 1989; Camerlenghi et al., 1992, 1995; Fig. 14). Further information has come from side-scan sonar mapping, combined with sampling (Galindo-Zaldivar et al., in press). Numerous other tectonic trends with a relief of up to 130 m, identified by Kenyon et al. (1982), may also be mud domes. Details of the Napoli Dome, in particular, were revealed by the site surveys conducted for Leg 160 during the TREDMAR-3 cruise (Limonov et al., 1994). Further details are given in the site chapters for Sites 970 and 971 (this volume).

One of the most remarkable recent discoveries during the TREDMAR-3 cruise (Limonov et al., 1994; Cita et al., 1994, in press) is that a number of mud domes can be seen to have been recently active. This is shown by the freshness of the extrusive features imaged on wide-beam and narrow-beam side-scan sonar records. Also, a video

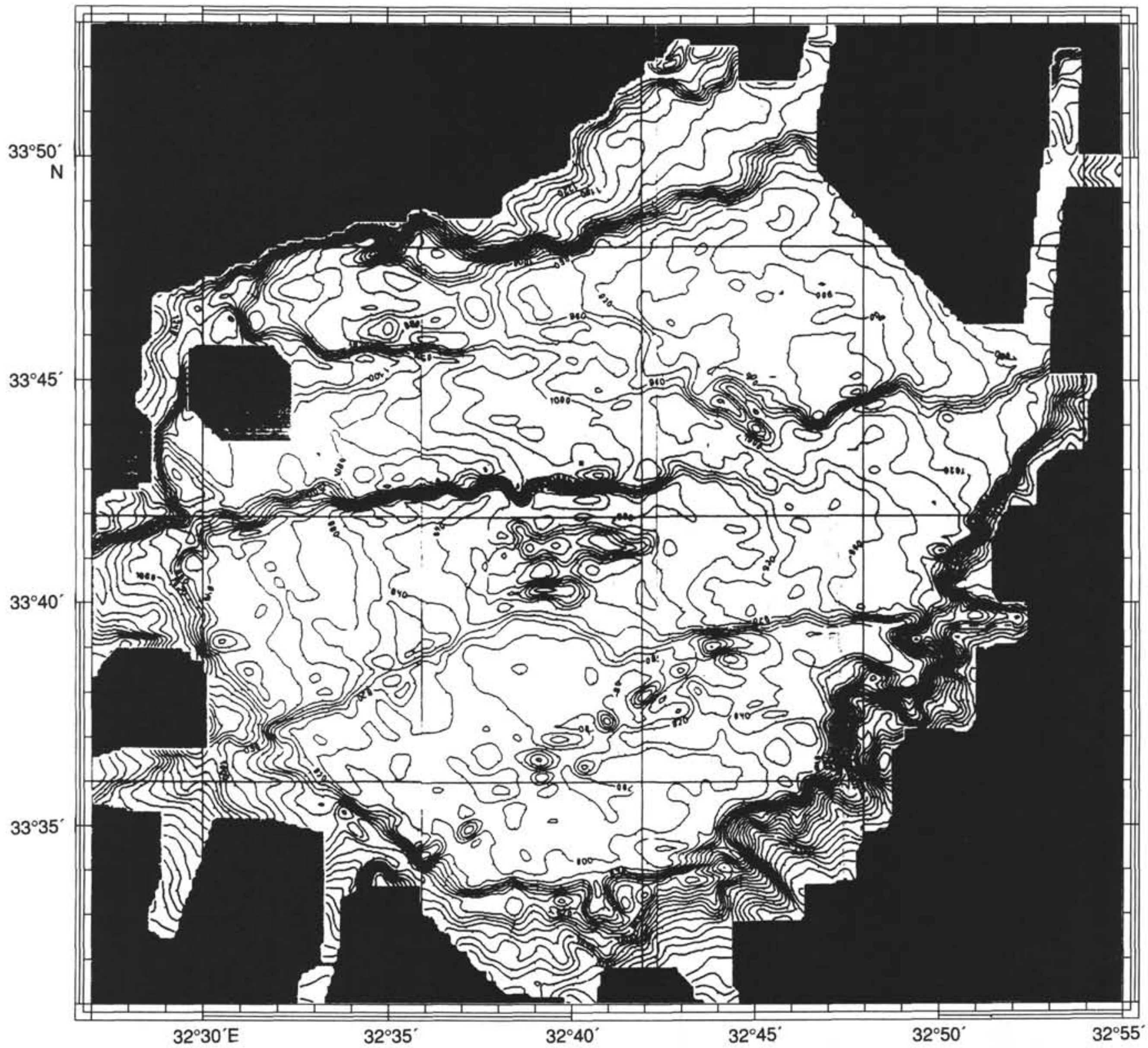

Figure 6. Swath bathymetry of the plateau area of the Eratosthenes Seamount. Note the nearly east-west lineaments, which are interpreted as an active fault zone. The data were collected during a cruise of *Academician Nikolai Strakhov*.

camera survey over the Napoli and Moscow Domes revealed numerous small holes of various shapes and sizes (up to 1 m across) that are interpreted as fluid-escape vents (Cita et al., 1994, in press). Such holes are surrounded by varicolored crusts that could be bacterial mats. Large numbers of shells are also present near the inferred vents, together with tracks interpreted as traces of unknown organisms. Clam shells belonging to the families *Vesicomyidae* and *Lucinidae* were sampled on Napoli Dome, and they indicate the existence of fauna linked to chemosynthetic processes associated with venting (Corselli and Bossi, in press).

Cores taken from the Napoli Dome contain highly disturbed sediment referred to by the term "mud breccia" (Cita et al., 1989; Staffini et al., 1993), which is overlain by a veneer of Holocene ooze. The disturbed sediment becomes finer grained upward and passes transitionally into pelagic marls. The actual boundary between the Holocene ooze and the underlying disturbed sediment is marked by an oxidized interval, up to 10 cm thick. The highly disturbed sediment comprises mainly a gray clay and silt-sized matrix that supports centimeter-sized subrounded clasts of semi-indurated sediment. Rounded holes up to 5 mm in diameter are visible in many of the split core sections, giving rise to a mousselike texture, inferred to result from gas expansion. Also, the cores had an odor of hydrogen sulfide. In some cases, extrusion of the core from the core liner of a whole-round sample of the mud breccia produced instantaneous expansion of the sediment with destruction of the sediment fabric.

It is worth noting that some readers might find the term "mud breccia" confusing, because to most geologists, this implies a sedimentary deposit composed of angular fragments that are mainly clast supported with only subordinate matrix. Also, the term mud breccia was coined in the context of the mud diapir hypothesis. In contrast, in the mud volcano hypothesis, the mud breccia deposits would more readily be interpreted as debris flows. Debris flows comprise clasts that commonly range from angular to subrounded in a matrix that is commonly dominated by mud or silt. The material cored at Sites 970

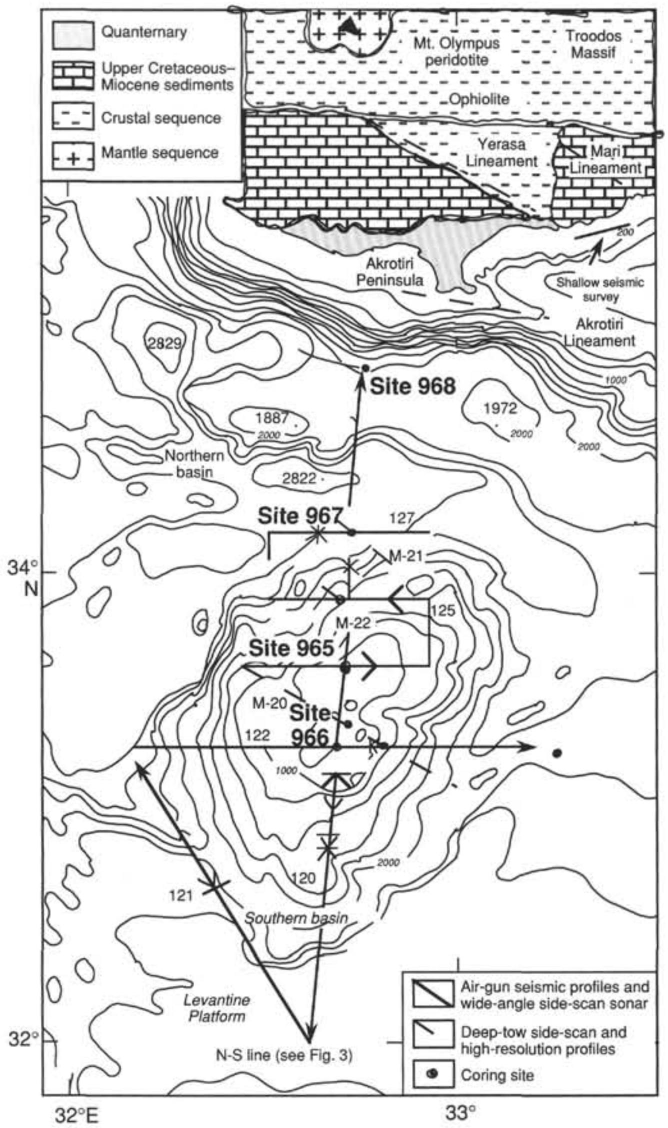

Figure 7. Location of the sites drilled during Leg 160 between the Eratosthenes Seamount and Cyprus. Note that the adjacent onshore geology of southern Cyprus is dominated by the Troodos ophiolite and the Tertiary carbonate section. From Robertson et al. (in press).

and 971 ranges greatly in consistency. The most common lithology is a sandy mudstone with scattered, mainly subangular clasts of variable lithology. The sediments are almost entirely matrix supported and were identified as debris flows.

Clasts that were cored prior to Leg 160 are typically gray, set in a brown matrix of hemipelagic-type sediment. The smallest clasts (millimeter sized) are rounded and indurated, whereas the larger clasts (centimeter sized) are commonly subangular and less indurated. The clasts vary in lithology from silty marls (most common) to scaly siltstones, indurated quartzose sandstones, and biocalcarenite. The carbonate fraction of the breccia from the Olimpi mud dome field ranges from 12% to 34%, with lower concentrations typically found in finer grained sediments. The natural water content varies from 33% to 51%, compared with 51% in the upper hemipelagic sediment, and is slightly higher than the saturation water content.

Cores from the slopes of the mud domes show an increased thickness of hemipelagic sediment downslope, with evidence of sediments that are highly disturbed by local hiatuses, graded beds, possible microfaulting, and gas-escape structures. Carbonate crusts were found mostly in slope sections within hemipelagic sediments.

The matrix of the mud breccias is assumed to be mostly late Aptian to Albian in age based on calcareous plankton analyses (Premoli Silva et al., in press). In addition, calcareous nannofossils and planktonic foraminifers of Late Cretaceous and early Miocene age were found locally. Within the debris flows of the Olimpi Field, the nannofossil and foraminifer content indicates that mixed fauna of mainly Oligocene and middle Miocene species are present (Staffini et al., 1993). The clasts contain nannofossils that mostly indicate Miocene and Pliocene ages (with Oligocene and Cretaceous forms also locally recognized); the foraminifers are mainly of middle Miocene age, with the Oligocene forms common only in the Pan di Zucchero Dome. Sporadic pre-Oligocene (i.e., Eocene and Cretaceous) faunas in the debris flows are probably reworked.

Microfossil assemblages also provide the best evidence for the age and timing of activity of the mud domes. For example, slope cores from the Napoli mud dome indicate that normal sedimentation resumed just prior to sapropel S1 deposition (approximately 10 ka). Cores from the crest of the dome do not exhibit S1, and it is thus possible that some activity persisted longer on the crest than on the flanks of the dome before being blanketed by hemipelagic sediment. Such arguments, however, depend on assumptions of relatively constant background sedimentation over the domes, which may not be valid. In addition, nannofossils that are assumed to have been derived from the mud dome are seen in surrounding hemipelagic sediments, ranging in age from 300,000 yr to the time of sapropel S6 deposition. This material was perhaps redeposited by low-density gravity flows, and, thus, suggests a prolonged period of upbuilding of the mud domes. Evidence of activity also comes from the pretephra Y-5 sediments (30,000–40,000 ka), and from the presence of sediments assumed to have been reworked from the Napoli Dome between sapropels S5 and S4 (~125 and ~100,000 ka) (Camerlenghi et al., 1992, 1995). However, a striking finding of Leg 160 is that the Napoli Dome is older than 1 Ma.

The provenance of the clasts in the mud domes could be interpreted by comparison with the sediment on the African plate approaching the active margin. These sediments, on the Sirte Abyssal Plain, are inferred based on seismic evidence to be located between 1.5 and 3.2 km beneath the seafloor, assuming that known rates of sedimentation (determined from piston cores) remained constant during the Tertiary. However, a deeper origin for the mud can be inferred from the presence of hydrocarbon gases, notably methane with a rather light carbon isotopic composition and appreciable quantities of C_{2+} hydrocarbon gases present in small percentages (1%–3%), which suggest a partly thermogenic, and thus deep, origin (Camerlenghi et al., 1995).

The taper of the Mediterranean Ridge dips at 22° to the north, and if this is taken as the angle of subduction and projected approximately 100 km northward to the mud volcano sites, then the depth of origin of the extruded mud could lie from 5.3 to 7.0 km (Camerlenghi et al., 1995).

Away from the mud domes, typical Holocene and Pleistocene sediments are composed of calcareous nannofossil ooze with foraminifers and a minor terrigenous fraction. The carbonate content ranges from 60% to 100%. The terrigenous fraction of the sediment includes clays, quartz, plagioclase, and potassium feldspar, derived mainly from the Nile and the Sahara region (Camerlenghi et al., 1992, 1995).

The actual geometry of the mud domes and the processes by which they are extruded are presently under debate. The relatively detailed OKEAN side-scan sonar survey conducted by TREDMAR-3 (Limonov et al., 1994) revealed synclinal rims around the crestal areas that were interpreted as vents of true mud volcanoes. Semicircular patterns of lineations on the flanks of the domes were thought to represent debris flows that erupted from the tops of mud volcanoes. On the other hand, Camerlenghi et al. (1992, 1995) noted that the sediment reflectors around the mud domes dip toward the conduit. They

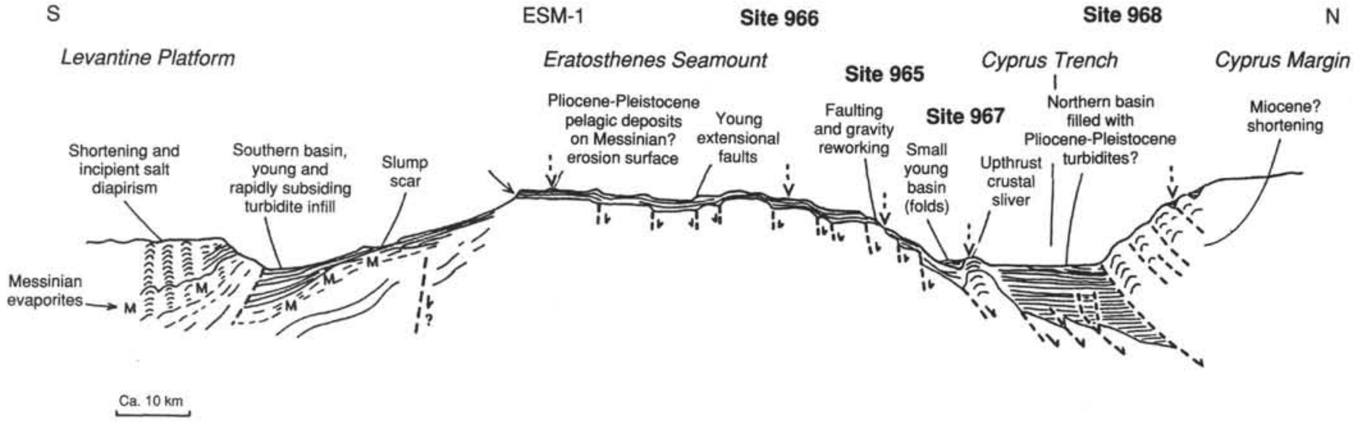

Figure 8. Tectonic interpretation of seismic reflection data collected during the site surveys for Leg 160. The Eratosthenes Seamount is envisaged as breaking up and being thrust northward beneath the Cyprus active margin. Drilling during Leg 160 confirmed this interpretation and, in addition, provided key evidence for the timing of events and the nature of the lithology within the seamount. From Robertson et al. (1994, in press) and Limonov et al. (1994).

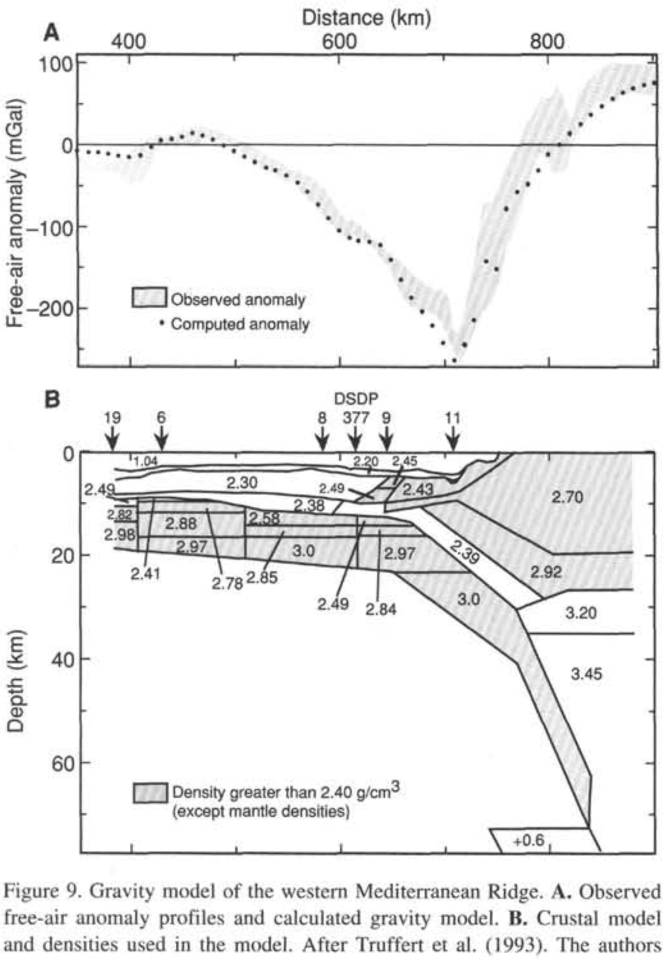

Figure 9. Gravity model of the western Mediterranean Ridge. **A.** Observed free-air anomaly profiles and calculated gravity model. **B.** Crustal model and densities used in the model. After Truffert et al. (1993). The authors considered that the inner unit of the Mediterranean Ridge (density 2.43 to 2.70 g/cm^3) acted as a backstop, which may be relevant to the formation of mud volcanoes in this area.

suggested that concentric seafloor collapse had taken place around individual mud domes. Similar features are reported from mud volcanos in the Barbados Ridge (Brown and Westbrook, 1988). The diameter of the inferred depressions is in some cases larger than the mud domes themselves. Larger depressions appear to be only partly filled with unusually low-viscosity muds of relatively high fluid content. A model can thus be inferred, by analogy with terrestrial magmatic volcanoes, in which initial extrusion is followed by collapse to form a large caldera that is then filled with initially fluid material; conical structures of more plastic material then form prior to extinction. A possible cause of such large-scale early collapse could be rapid gas release (Camerlenghi et al., 1995).

The relatively plastic (rather than fully liquified) rheology of the pebbly muds, combined with the evidence in cores of local frictional deformation of some individual pebbles, was in the past taken as evidence that the mud was not merely extruded as a consequence of ambient lithostatic load, but instead, was relatively forcefully ejected, perhaps driven by overpressuring of pore fluids and expansion of gases during upward extrusion. The role of evaporites in mud dome genesis was unclear. The presence of evaporite is documented on the entire Mediterranean Ridge by the occurrence of the "M" horizon, a strong seismic reflector (or multiple reflector) that commonly constitutes acoustic basement on high-resolution, single-channel seismic profiles and is attributed to the base of the Pliocene. However, the thickness and composition of any evaporites beneath the mud domes is poorly constrained. One possibility is that subduction-accretion effects may have led to the buildup of pore pressures beneath impermeable evaporites, followed by puncturing and the extrusion of mud. New data from the interstitial fluids of the mud domes collected during Leg 160 have shed additional light on the possible role of evaporites at depth. Available evidence suggests that the mud volcanoes are developed in areas where evaporites may be relatively thin (i.e., approximately 100 m), specifically along the inner (i.e., northerly) escarpment of the crestal plateau of the Mediterranean Ridge. This is an area where the accretionary wedge apparently is in the process of being backthrust against a backstop, composed of the Inner Plateau area and the Hellenic Trench (which may be a related flexural feature—Figs. 14, 15). One possibility is that seaward-dipping backthrusts provided long-lived zones of egress for overpressured fluids trapped beneath thick (i.e., 1000 m) evaporites within the accretionary complex to the south (Fig. 16).

In summary, the drilling of mud domes during Leg 160 was designed to shed light on their origin on the Mediterranean Ridge. Such mud domes have never been drilled in deep water before and, thus, represent a challenge for the Ocean Drilling Program. Similar mud domes are widespread in tectonically active areas of the Mediterranean and elsewhere and may represent significant sources of natural pollutants within the Mediterranean (e.g., hydrocarbons). A better

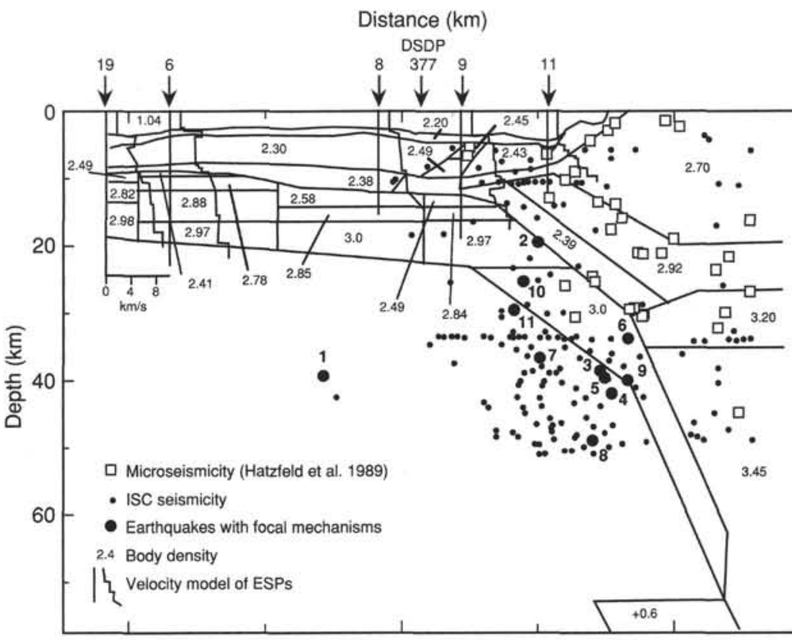

Figure 10. Summary of recent expanding spread profile (ESP) data across the western Mediterranean Ridge. Arrows mark ESP locations, and velocity models are shown for selected ESPs. Squares indicate the distribution of microseismicity and small dots indicate the microseismic distribution from the International Seismic Centre (ISC) catalogue. The large dots numbered from 1 to 11 represent the location of seismic events for which focal mechanisms have been determined. DSDP Site 377 is located on this cross section. After Truffert et al. (1993). The mud volcano sites drilled during Leg 160 are located above the point of inflection in the downgoing slab.

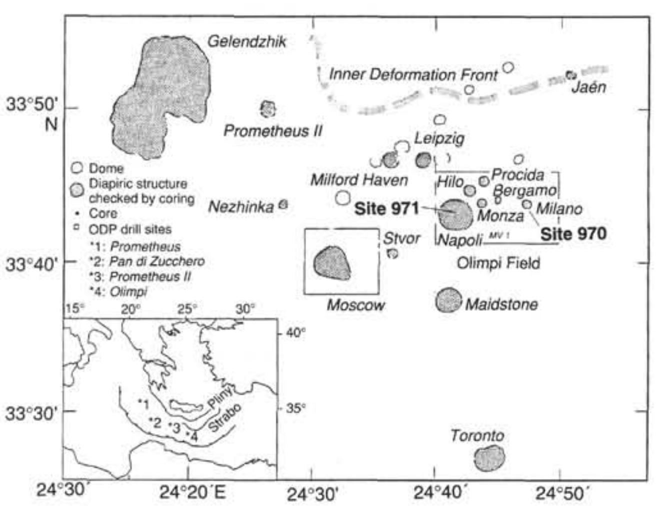

Figure 11. Location of the known mud volcanoes surveyed during the preparatory work for Leg 160. Note the location of the large Napoli Dome (Site 970) and the smaller Milano Dome (Site 971). After Limonov et al. (1994).

understanding of fluid processes associated with the formation of mud domes is therefore also relevant to determining the long-term effects of pollution of the deep-sea environment. It was intended that the results would test alternative models for mud dome origin, notably the mud diapir vs. mud volcano hypotheses, and also shed light on the conditions of extrusion and ultimate origin at depth.

ACCRETIONARY AND COLLISIONAL PROCESSES

Collisional and accretionary processes in the westernmost part of the eastern Mediterranean Sea were also studied during Leg 160, although safety and time considerations limited penetration. In addition, Site 964 was scheduled for drilling on the basis of its paleoceanography, but included aspects of tectonic interest.

One aspect was the study of the toes of accretionary wedges, including the Pisano Plateau, and the rugged seafloor adjacent to the deformation front in the Ionian Sea (Site 964). A more advanced collisional setting was cored at Site 963, south of Sicily, as part of another paleoceanographic study. Following drilling of the other objectives, there was insufficient time to drill all three of the Ionian transect sites as indicated in the "Leg 160 Scientific Prospectus." The site that was drilled lower on the deformation front (Site 972) went only to 100 m below seafloor (mbsf) because of technical problems. However, drilling at Site 973, which is slightly higher on the deformation front, reached 200 mbsf and produced useful tectonic-related information.

The Ionian deformation front represents one segment of the feather edge of the Mediterranean Ridge accretionary wedge. The deformation front (Fig. 1) is defined by a sudden change from flat and layered reflections on the Ionian Abyssal Plain to hyperbolic patterns produced by the cobblestone topography of the Mediterranean Ridge. A 10.5-km-long, 310-m-high "seahill" oriented in a southwest-northeast direction (about 45° with respect to the regional trend of the Mediterranean Ridge) is located a few kilometers seaward of the deformation front (Hieke, 1978). The highest part of the Victor Hensen structure can be traced in the sub-bottom for more than 60 km in a southwest-northeast direction. Recently acquired multichannel-seismic lines (1992 cruise *Valdivia* 120 MEDRAC; W. Heike, pers. comm., 1994) provide evidence of several elongate structures that indicate intense pre-Messinian tectonic activity under the Ionian Abyssal Plain, seaward of the Ionian deformation front. The structures are elevated from a sequence of partly tilted pre-Messinian sediments. However, the tops of some elongate structures are free of evaporites. The style of deformation is considered indicative of extensional block faulting. Because the Messinian evaporites and overlying Pliocene-Pleistocene turbidites show less evidence of deformation, the main part of the tectonic phase that produced these structures must be, at the latest, of Messinian age (about 5 Ma). However, deformation continued during the Pliocene-Pleistocene with synsedimentary extensional faulting (Avedik and Hieke, 1981; Hieke and Wanninger, 1985; Hirschleber et al., 1994). The Victor Hensen structure interacts with the present-day deformation front, indicating that a complicated pattern of varying lithologies and thicknesses of sediments has become incorporated in the Mediterranean Ridge accretionary complex. The styles of initial deformation are, thus, expected to be correspondingly different. The proposed transect of shallow holes was intended to sample the incoming sediment section overly-

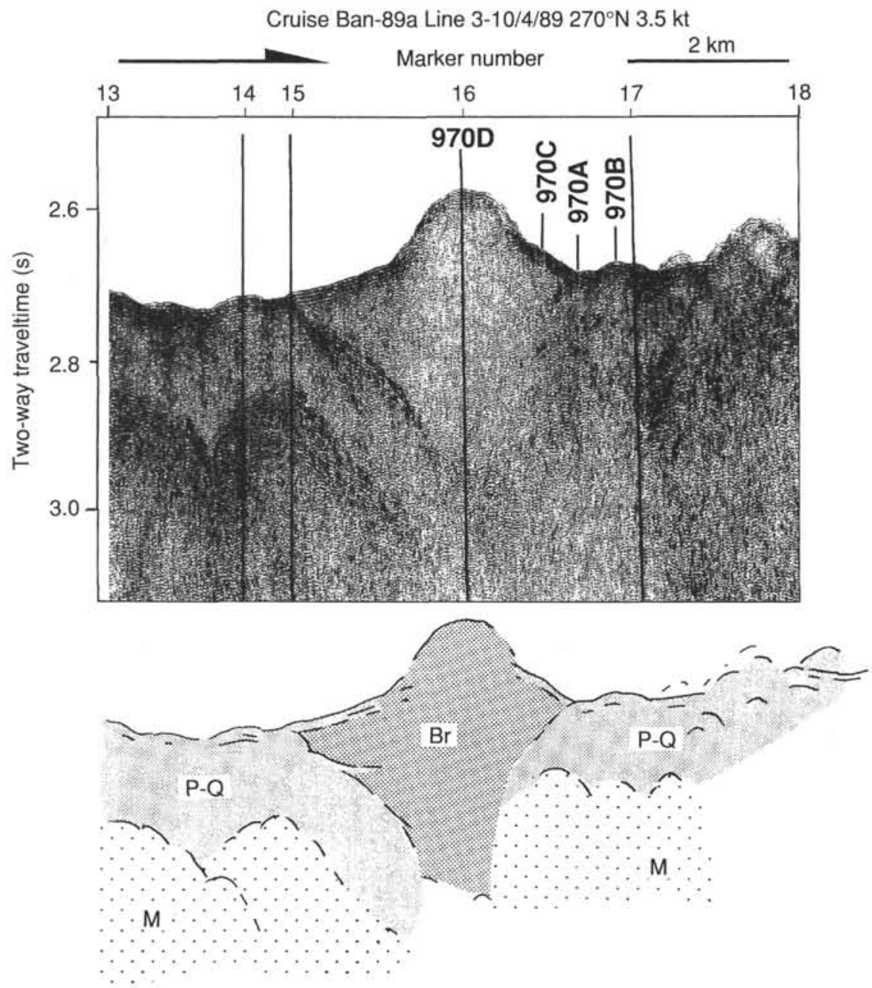

Figure 12. Setting of the Milano Dome inferred prior to Leg 160. Note the location of Site 970 on the crest and flanks of the Milano mud volcano on the single-channel sparker line shown in the upper part of the figure. The interpretation of the seismic line in the lower part of the figure shows the inferred diapir (Br) cutting across the interpreted Miocene and overlying Pliocene-Pleistocene units. After Camerlenghi et al., 1995).

ing the buried Victor Hensen structure in the abyssal plain and the post-Messinian deformed sediment of the ridge. It could also reveal some evidence of the nature and timing of accretionary processes along the Inner Deformation Front, notably the possible role of Messinian evaporites, as determined from interstitial-fluid composition.

Although the sites were selected to improve understanding of the origin of sapropels, information of tectonic interest was also obtained. In the area of the Pisano Plateau, drilled at Site 964, the floor of the Ionian Abyssal Plain is being underthrust beneath the Calabrian arc. This results in the detachment of deep-sea sediments to form an accretionary wedge (Finetti, 1976, 1982). The location of the site, elevated approximately 200 m above the floor of the Ionian Abyssal Plain, implies a location near the toe of the Calabrian accretionary wedge (Kastens, 1984; Morelli et al., 1975). The site thus provided some information on the sedimentation and tectonic history of an accretionary setting located adjacent to the Ionian deformation front.

Another setting of tectonic interest was drilled at Site 963 on the foreland of North Africa within the collision zone with the overriding Eurasian plate. This part of North Africa formed a promontory that jutted into the Tethyan ocean to the north during Mesozoic–early Tertiary time. This promontory is currently in the process of colliding with the overriding plate and its tectonic setting is as follows:

During the Late Cretaceous?–early Tertiary, Tethyan oceanic crust between Africa and Eurasia was subducted northward, building up accretionary wedges that migrated southward through time (for a recent overview, see Robertson and Grasso, 1995). Subduction was accompanied by the rifting of Corsica and Sardinia from the southern margin of Eurasia, in Provence, accompanied by opening of the North Balearic Basin during the late Oligocene–early Miocene. This was followed by collision with Apulia and thrusting in the Northern Apennines. Shortening affected Sicily by the late Oligocene–Langhian, when a series of thrust sheets was emplaced onto the former North African shelf edge, giving rise to southward-migrating foreland basins, with the deposition of marine siliciclastic sediments derived from North Africa (i.e., Numidian Flysch; Compagnoni et al., 1989). Associated clastic sediments (e.g., Capo d'Orlando Flysch) accumulated in perched basins above thrust sheets of emplaced crystalline rocks (i.e., Calabride units).

Convergence continued during the Miocene. In more southerly areas of what is now central southern Sicily, by the end of the Burdigalian, large successor foreland basins were developing ahead of advancing thrust sheets (i.e., Maghrebian units). By the late Serravallian–Tortonian, the earlier foreland basins were deformed by renewed thrusting. During the late Tortonian, proximal deltaic sediments were deposited on recently deformed areas, and the foreland basin then shifted southward to its present position (i.e., Caltanisetta Basin). The orogenic units (i.e., Maghrebian thrust sheets) can be traced from Sicily under the sea to northern Tunisia (Ben Avraham et al., 1992). Marine studies of submerged parts of the Maghrebian chain reveal an imbricate thrust stack consisting of European- and North African-derived units and lower Miocene syntectonic units (Catalano et al., 1987, 1993a, 1993b).

During the Pliocene-Pleistocene, the area was dominated by thin-skinned thrust tectonics, with the development of a duplex structure

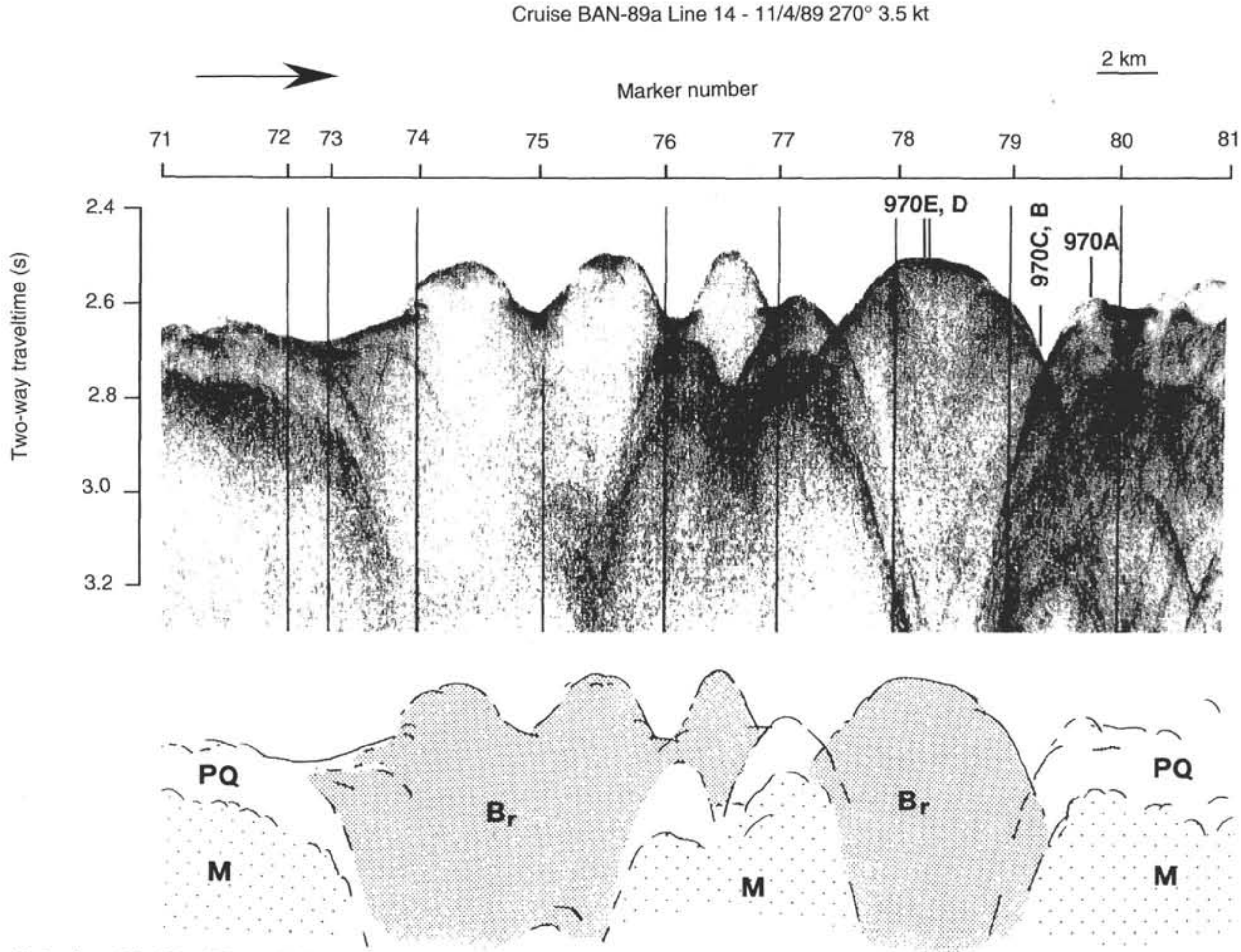

Figure 13. Setting of the Napoli Dome inferred prior to Leg 160. Note the location of Site 971 on the crest and flanks of the Napoli mud volcano on the single-channel sparker line shown in the upper part of the figure. The interpretation of the seismic line in the lower part of the figure shows the inferred diapirs (Br) cutting across the interpreted Miocene and overlying Pliocene-Pleistocene units. After Camerlenghi et al. (1995).

(Argnani, 1989; Argnani et al., 1989; Trincardi and Argnani, 1990). The irregularly shaped continental margin of North Africa (i.e., Hyblean Plateau and Adventure Plateau) controlled the collision with the advancing thrust front. Subduction, however, continued farther northeast, along the Calabrian arc, giving rise to continued extension within the Tyrrhenian Sea, as documented during Leg 107 (Kastens, Mascle, et al., 1990; Kastens et al., 1988).

Tectonic lineaments on the North African continental margin were reactivated during the Pliocene-Pleistocene, including the Pantelleria rift system in the Sicily Platform and the north-south axis of adjacent Tunisia (Gardiner et al., in press). The Pantelleria rift system comprises three main grabens, the Malta, Pantelleria, and Linosa basins, which are linked by wrench faults (Catalano et al., 1987, 1993a, 1993b). The basins trend almost at right angles to the compressional front located to the north. Rifting apparently began in the late Miocene, based on sedimentary evidence from adjacent areas (e.g., Maltese islands). The Pantelleria rift system probably developed to accommodate the effects of oblique collision with the irregularly shaped North African continental margin (Jongsma et al., 1985, 1987).

In conclusion, the tectonic component of Leg 160 was conceived as a contribution to the understanding of incipient collision-related processes in the Eastern Mediterranean, with a major focus on the tectonic history of the Eratosthenes Seamount and the mud volcanoes of the Mediterranean Ridge. The aim in each case was to shed light on fundamental processes that operate on a global basis.

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REFERENCES

Argnani, A., 1989. The Gela Nappe: evidence of accretionary melange in the Maghrebian foredeep of Sicily. *Mem. Soc. Geol. Ital.,* 38:419–428.

Argnani, A., Cornini, S., Torelli, L., and Zitellini, N., 1989. Diachronous foredeep system in the Neogene-Quaternary of the Strait of Sicily. *Mem. Soc. Geol. Ital.,* 38:407–417.

Avedik, F., and Hieke, W., 1981. Reflection seismic profiles from the Central Ionian Sea (Mediterranean) and their geodynamic interpretation. *"Meteor" Forschungsergeb. Reihe C,* 34:49–64.

Ben-Avraham, Z., Boccaletti, M., Grasso, M., Lentini, F., Torelli, L., and Tortorici, L., 1992. Principi domini strutturali originatisi dalla collisione continentale neogenico-quaternaria del Mediterraneo centrale. *Mem. Soc. Geol. Ital.,* 45:453–462.

Ben-Avraham, Z., and Nur, A., 1986. Collisional processes in the Eastern Mediterranean. *Geol. Rundsch.,* 75:209–217.

Ben-Avraham, Z., Shoham, Y., and Ginzburg, A., 1976. Magnetic anomalies in the eastern Mediterranean and the tectonic setting of the Eratosthenes Seamount. *Geophys. J. R. Astron. Soc.,* 45:105–123.

Brown, K., and Westbrook, G.K., 1988. Mud diapirism and subcretion in the Barbados Ridge accretionary complex: the role of fluids in accretionary processes. *Tectonics,* 7:613–640.

Camerlenghi, A., Cita, M.B., Della Vedova, B., Fusi, N., Mirabile, L., and Pellis, G., 1995. Geophysical evidence of mud diapirism on the Mediterranean Ridge accretionary complex. *Mar. Geophys. Res.,* 17:115–141.

Camerlenghi, A., Cita, M.B., Hieke, W., and Ricchiuto, T.S., 1992. Geological evidence for mud diapirism on the Mediterranean Ridge accretionary complex. *Earth Planet. Sci. Lett.,* 109:493–504.

Catalano, R., D'Argenio, B., and Torelli, L., 1987. A geologic section from Sardinia to Sicily Straits based on seismic and field data. *In* Boriani, A., et al. (Eds.), *Lithosphere in Italy: Advances in Earth Science Research:* Roma (Academica Nazionale die Lincei), 109–125.

Catalano, R., Di Stefano, P., Nigro, F., and Vitale, F.P., 1993a. Sicily mainland and its offshore: a structural comparison. *In* Max, M.D., and Colantoni, P. (Eds.), *Geological Development of the Sicilian-Tunisian Platform.* UNESCO Rep. Mar. Sci., 58:19–24.

Catalano, R., Infuso, S., and Sulli, A., 1993b. The Pelagian Foreland and its northward foredeep: Plio-Pleistocene structural evolution. *In* Max, M.D., and Colantoni, P. (Eds.), *Geological Development of the Sicilian-Tunisian Platform.* UNESCO Rep. Mar. Sci., 58:37–42.

Cita, M.B., Camerlenghi, A., Erba, E., McCoy, F.W., Castradori, D., Cazzani, A., Guasti, G., Giambastiani, M., Lucchi, R., Nolli, V., Pezzi, G., Redaelli, M., Rizzi, E., Torricelli, S., and Violanti, D., 1989. Discovery of mud diapirism in the Mediterranean Ridge: a preliminary report. *Boll. Soc. Geol. Ital.,* 108:537–543.

————, 1994. Fluid venting, mud volcanoes and mud diapirs in the Mediterranean Ridge. *Rend. Fis. Acc. Lin. Roma,* 5:159–169.

Cita, M.B., Woodside, J.M., Ivanov, M.K., Kidd., R.B., Limonov, A.F., et al., in press. Fluid venting from a mud volcano in the Mediterranean Ridge Diapiric Belt. *Terra Nova.*

Compagnoni, R., Morlotti, E., and Torelli, L., 1989. Crystalline and sedimentary rocks from the scarps of the Sicily-Sardinias Trough and Cornaglia Terrace (southwestern Tyrrhenian Sea, Italy): paleogeographic and geodynamic implications. *In* Beccaluva, L. (Ed.), *Ophiolites and Lithosphere of Marginal Seas.* Chem. Geol., 77:375–398.

Corselli, C., and Bossi, D., in press. First evidence of benthic communities based on chemosynthesis on the Napoli mud volcano (Eastern Mediterranean). *Mar. Geol.*

de Voogd, B., Truffert, C., Chamot-Rooke, N., Huchon, P., Lallemant, S., and Le Pichon, X., 1992. Two-ship deep seismic soundings in the basins of the eastern Mediterranean Sea (Pasiphae cruise). *Geophys. J. Int.,* 109:536–552.

Dewey, J.-F., 1980. Episodicity, sequence, and style at convergent plate boundaries. *In* Strangeway, D.W. (Ed.), *The Continental Crust and Its Mineral Deposits.* Spec. Pap.—Geol. Assoc. Can., 20:553–573.

Eaton, S., and Robertson, A.H.F., 1993. The Miocene Pakhna Formation, Cyprus, and its relationship to the Neogene tectonic evolution of the Eastern Mediterranean. *Sediment. Geol.* 86:273–296.

Finetti, I., 1976. Mediterranean Ridge: a young submerged chain associated with the Hellenic Arc. *Boll. Geofis. Teor. Appl.,* 19:31–65.

————, 1982. Structure, stratigraphy and evolution of central Mediterranean. *Boll. Geofis. Teor. Appl.,* 24:247–315.

Fytikas, M., Innocenti, P., Manetti, P., Mazzuoli, R., Peccerillo, A., and Villari, L., 1984. Tertiary to Quaternary evolution of volcanism in the Aegean region. *In* Dixon, J.E., and Robertson, A.H.F. (Eds.), *The Geological Evolution of the Eastern Mediterranean.* Spec. Publ. Geol. Soc. London, 17:687–700.

Galindo-Zaldivar, J., Nieto, L., Woodside, J.M. and Limonov, A.F., in press. Structural features of mud volcanoes and the fold system of the Mediterranean Ridge. *Mar. Geol.*

Gardiner, W., Grasso, M., and Sedgeley, D., in press. Plio-Pleistocene fault movement as evidence for mega-block kinematics within the Hyblean-Malta Plateau, central Mediterranean. *J. Geodyn.*

Gass, I.G., and Masson-Smith, D., 1963. The geology and gravity anomalies of the Troodos Massif, Cyprus. *Phil. Trans. R. Soc. London A,* 255:417–467.

Giese, P., Nicolich, R., and Reutter, K.J., 1982. Explosion seismic crustal studies in the Alpine-Mediterranean region and their implication to tectonic processes. *In* Berckhemer, H., and Hsü, K. (Eds.), *Alpine-Mediterranean Geodynamics.* AGU-GSA Geodyn. Ser., 6:39–74.

Hieke, W., 1978. The Victor Hensen Seahill: part of a tectonic structure in the Central Ionian Sea. *Mar. Geol.,* 26:M1–M5.

Hieke, W., and Wanninger, A., 1985. The Victor Hensen Seahill (Central Ionian Sea): morphology and structural aspects. *Mar. Geol.,* 64:343–350.

Horvath, F., and Berckhemer, H., 1982. Mediterranean back arc basins. *In* Berckhemer, H., and Hsü, K.J. (Eds.), *Alpine-Mediterranean Geodynamics.* Ber. Inst. Meteorol. Geophys. Univ. Frankfurt/Main, Geodyn. Ser., 7:141–173.

Hsü, K.J., Montadert, L., et al., 1978. *Init. Repts. DSDP,* 42 (Pt. 1): Washington (U.S. Govt. Printing Office).

Jongsma, D., van Hinte, J.E., and Woodside, J.M., 1985. Geologic structure and neotectonics of the north African continental margin, south Sicily. *Mar. Pet. Geol.,* 2:156–179.

Jongsma, D., Woodside, J.M., King, G.C.P., and van Hinte, J.E., 1987. The Medina Wrench: a key to the kinematics of central and eastern Mediterranean over the past 5 Ma. *Earth Planet. Sci. Lett.,* 82:87–106.

Kastens, K.A., 1984. Earthquakes as a triggering mechanism for debris flows and turbidites on Calabrian Ridge. *Mar. Geol.,* 55:13–33.

————, 1991. Rate of outward growth of the Mediterranean Ridge accretionary complex. *Tectonophysics,* 199:28–50.

Kastens, K.A., Breen, N., and Cita, M.B., 1992. Progressive deformation on an evaporite-bearing accretionary complex: SeaMARC 1, Seabeam, and piston-core observations from the Mediterranean Ridge. *Mar. Geophys. Res.,* 14:249–298.

Kastens, K.A., Mascle, J., et al., 1990. *Proc. ODP, Sci. Results,* 107: College Station, TX (Ocean Drilling Program).

Kastens, K., Mascle, J., Auroux, C.A., Bonatti, E., Broglia, C., Channell, J., Curzi, P., Emeis, K.-C., Glacon, G., Hasegawa, S., Hieke, W., Mascle, G., McCoy, F., McKenzie, J., Mueller, C., Rehault, J.P., Robertson, A., Sartori, R., Sprovieri, R., and Torii, M., 1988. ODP Leg 107 in the Tyrhennian Sea: insights into passive margin and back-arc basin evolution. *Geol. Soc. Am. Bull.,* 100:1140–1156.

Kelling, G., Gokcen, S.L., Floyd, P.A., and Gokcen, N., 1987. Neogene tectonics and plate convergence in the Eastern Mediterranean: new data from southern Turkey. *Geology,* 15:425–429.

Kempler, D., 1993. Tectonic patterns in the Eastern Mediterranean [Ph.D. thesis]. Hebrew Univ., Jerusalem.

Kempler, D., and Ben-Avraham, Z., 1987. The tectonic evolution of the Cyprean Arc. *Ann. Tecton.,* 1:58–71.

Kempler, D., and Garfunkel, Z., 1991. The northeastern Mediterranean triple junction from a plate tectonic kinematic point of view. *Bull. Tech. Univ. Istanbul,* 44:425–454.

————, 1992. The Eratosthenes Seamount: a fossil superstructure in the Eastern Mediterranean. *Rapp. Comm. Int. Mer Medit.,* 33:391.

Kenyon, N.H., Belderson, R.H., and Stride, A.H., 1982. Detailed tectonic trends on the central part of the Hellenic Outer Ridge in the Hellenic Trench System. *In* Leggett, J.K. (Ed.), *Trench-Forearc Geology.* Geol. Soc. Spec. Publ. London, 10:335–343.

Kissel, C., and Laj, C., 1988. The Tertiary geodynamical evolution of the Aegean Arc: a paleomagnetic reconstruction. *Tectonophysics,* 146:183–201.

Krasheninnikov, V.A., and Hall, J.K., 1994. *Geological Structure of the Northeastern Mediterranean (Cruise 5 of Ctr Research Vessel) "Acdemik Nikolas Strakhov":* Jerusalem (Historical Productions–Hall Ltd.).

Le Pichon, X., 1982. Land-locked oceanic basins and continental collision: the Eastern Mediterranean as a case example. *In* Hsü, K.J. (Ed.), *Mountain Building Processes:* New York (Academic), 201–211.

Le Pichon, X., and Angelier, J., 1979. The Hellenic arc and trench system: a key to the neotectonic evolution of the eastern Mediterranean area. *Tectonophysics,* 60:1–42.

Le Pichon, X., Augustithis, S.S., and Mascle, J., 1982a. *Geodynamics of the Hellenic Arc and Trench.* Spec. Iss., Tectonophysics, 86.

Le Pichon, X., Lyberis, N., Angelier, J., and Renard, V., 1982b. Strain distribution over the East Mediterranean Ridge: a synthesis incorporating new Sea Beam data. *Tectonophysics,* 86:243–274.

Limonov, A.F., Woodside, J.M., and Ivanov, M.K. (Eds.), 1994. *Mud Volcanism in the Mediterranean and Black Seas and Shallow Structure of the Eratosthenes Seamount.* Initial Results of the Geological and Geophysical Investigations during the Third "Training-through-Research" Cruise of the *R/V Gelendzhik* (June–July 1993). UNESCO Rep. Mar. Sci., 64.

Makris, J., Ben-Avraham, Z., Behle, A., Ginzburg, A., Gieze, P., Steinmetz, L., Whitmarsh, R.B., and Eleftheriou, S., 1983. Seismic refraction profiles between Cyprus and Israel and their interpretation. *Geophys. J. R. Astron. Soc.,* 75:575–591.

Makris, J., and Stobbe, C., 1984. Physical properties and state of the crust and upper mantle of the eastern Mediterranean Sea as deduced from geophysical data. *Mar. Geol.,* 55:347–363.

Mart, Y., 1994. Ptolemais Basin: the tectonic origin of a Senonian marine basin underneath the southeastern Mediterranean Sea. *Tectonophysics,* 234:5–18.

McKenzie, D.P., 1972. Active tectonics of the Mediterranean region. *Geophys. J. R. Astron. Soc.*, 30:109–185.

————, 1978. Active tectonism in the Alpine-Himalayan belt: the Aegean Sea and the surrounding regions (tectonics of the Aegean region). *Geophys. J. R. Astron. Soc.*, 55:217–254.

Meulenkamp, J.E., van der Zwaan, G.J., and van Wamel, W.A., 1994. On Late Miocene to Recent vertical movements in the Cretan segment of the Hellenic arc. *Tectonophysics,* 234:53–72.

Meulenkamp, J.E., Wortel, M.J.R., van Wamel, W.A., Spakman, W., and Hoogeruijn Strating, E., 1988. On the Hellenic subduction zone and the geodynamic evolution of Crete since the late middle Miocene. *Tectonophysics,* 146:203–215.

Montadert, L., Letouzey, J., and Mauffret, A., 1978. Messinian event: seismic evidence. *In* Hsü, K.J., Montadert, L., et al., *Init. Repts. DSDP,* 42 (Pt. 1): Washington (U.S. Govt. Printing Office), 1037–1050.

Moores, E.M., and Vine, F.J., 1971. The Troodos Massif, Cyprus and other ophiolites as oceanic crust: evaluation and implications. *Philos. Trans. R. Soc. London A,* 268:443–466.

Morelli, C., Gantar, G., and Pisani, M., 1975. Bathymetry, gravity and magnetism in the Strait of Sicily and in the Ionian Sea (plus map sheets). *Boll. Geofis. Teor. Appl.,* 17:39–58.

Payne, A.S., and Robertson, A.H.F., in press. Neogene supra-subduction extension in the Polis graben system, west Cyprus. *J. Geol. Soc. London.*

Poole, A.J., and Robertson, A.H.F., 1992. Quaternary uplift and sea-level change at an active plate boundary, Cyprus. *J. Geol. Soc. London,* 148:909–921.

Premoli Silva, I., Erba, E., Spezzaferri, S., and Cita, M.B., in press. Variation in the age of the diapiric mud breccia along and across the axis of the Mediterranean Ridge accretionary complex. *Mar. Geol.*

Rabinowitz, P.D., and Ryan, W.B.F., 1970. Gravity anomalies and crustal shortening in the eastern Mediterranean. *Tectonophysics,* 10:585–608.

Robertson, A.H.F., 1977. Tertiary uplift history of the Troodos Massif, Cyprus. *Geol. Soc. Am. Bull.,* 88:1763–1772.

————, 1990. Tectonic evolution of Cyprus. *In* Malpas, J., Moores, E.M., Panayiotou, A., and Xenophontos, C. (Eds.), *Ophiolites: Oceanic Crustal Analogues: Proc. Symp. "Troodos 1987":* Nicosia, Cyprus (Minist. Agric. Nat. Resour.), 235–250.

————, 1992. Drilling the Eratosthenes Seamount: Mediterranean collision tectonics and Plio-Quarternary palaeo-oceanography in the light of the geology of Cyprus. *Rapp. Comm. Mer Medit.,* 33:393.

————, 1994. Role of the tectonic facies concept in orogenic analysis and its application to Tethys in the Eastern Mediterranean region. *Earth-Sci. Rev.,* 37:139–213.

Robertson, A.H.F., Eaton, S., Follows, E.J., and McCallum, J.E., 1991. The role of local tectonics versus global sea-level change in the Neogene evolution of the Cyprus active margin. *Spec. Publ. Int. Assoc. Sediment.,* 12:331–369.

Robertson, A.H.F., and Grasso, M., 1995. Overview of the Late Tertiary–Recent tectonic and palaeo-environmental development of the Mediterranean region. *Terra Nova,* 7:114–127.

Robertson, A.H.F., Kidd, R.B., Ivanov, M.K., and Limonov, A.F., in press. Eratosthenes Seamount, easternmost Mediterranean: evidence of active collapse and thrusting beneath Cyprus. *Terra Nova.*

Robertson, A.H.F., Kidd, R.B., Ivanov, M.K., Limonov, A.F., Woodside, J.M., Galindo-Zaldivar, J., and Nieto, L., 1994. Probing continental collision in the Mediterranean Sea. *Eos,* 75:233, 239.

Robertson, A.H.F., and Xenophontos, C., 1993. Development of concepts concerning the Troodos ophiolite and adjacent units in Cyprus. *In* Prichard, H.M., Alabaster, T., Harris, N.B. and Neary, C.R. (Eds), *Magmatic Processes and Plate Tectonics.* Geol. Soc. Spec. Publ. London, 70:85–120.

Ross, D.A., and Uchupi, E., 1977. Structure and sedimentary history of southeastern Mediterranean Sea—Nile Cone. *AAPG Bull.,* 61:872–902.

Rotstein, Y., and Kafka, A.L., 1982. Seismotectonics of the Cyprean Arc, Eastern Mediterranean Region: subduction, collision and arc jumping. *J. Geophys. Res.,* 87:7694–7707.

Ryan, W.B.F., Hsü, K.J., et al., 1973. *Init. Repts. DSDP,* 13 (Pt. 1): Washington (U.S. Govt. Printing Office).

Sengör, A.M.C., Görür, N., and Sarioلü, F., 1985. Strike-slip faulting and related basin formation in zones of tectonic escape: Turkey as a case study. *In* Biddle, K.T., and Christie-Blick, N. (Eds.), *Strike-slip Deformation, Basin Formation and Sedimentation.* Soc. Econ. Paleont. Mineral., Spec. Publ., 37:227–264.

Staffini, F., Spezzaferri, S., and Aghib, F., 1993. Mud diapirs of the Mediterranean Ridge: sedimentological and micropalaeontological study of the mud breccia. *Riv. Ital. Paleont. Stratigr.,* 99:225–254.

Trincardi, F., and Argnani, A., 1990. Gela submarine slide: a major basin-wide event in the Plio-Quaternary foredeep of Sicily. *Geo-Mar. Lett.,* 10:13–21.

Truffert, C., Chamot-Rooke, N., Lallemant, S., de Voogd, B., Huchon, P., and Le Pichon, X., 1993. The crust of the Western Mediterranean Ridge from deep seismic data and gravity modelling. *Geophys. J. Int.,* 114:360–372.

Underhill, J.R., 1989. Late Cenozoic deformation of the Hellenide foreland, western Greece. *Geol. Soc. Am. Bull.,* 101:613–634.

Woodside, J., and Bowin, C., 1970. Gravity anomalies and inferred crustal structure in the eastern Mediterranean Sea. *Geol. Soc. Am. Bull.,* 81:1107–1122.

Woodside, J.M., 1976. Regional vertical tectonics in the eastern Mediterranean. *Geophys. J. R. Astron. Soc.,* 47:493–514.

————, 1977. Tectonic elements and crust of the eastern Mediterranean Sea. *Mar. Geophys. Res.,* 317–354.

————, 1991. Disruption of the African Plate margin in the eastern Mediterranean. *In* Salem, M.J., Sbeta, A.M., Bakbak, M., and Rida, M. (Eds.), *The Geology of Libya:* Symp. Geol. Libya 3, 6:2319–2329.

Wortel, M.J.R., Goes, S.D.B., and Spakman, W., 1990. Structure and seismicity of the Aegean subduction zone. *Terra Nova,* 2:462–554.

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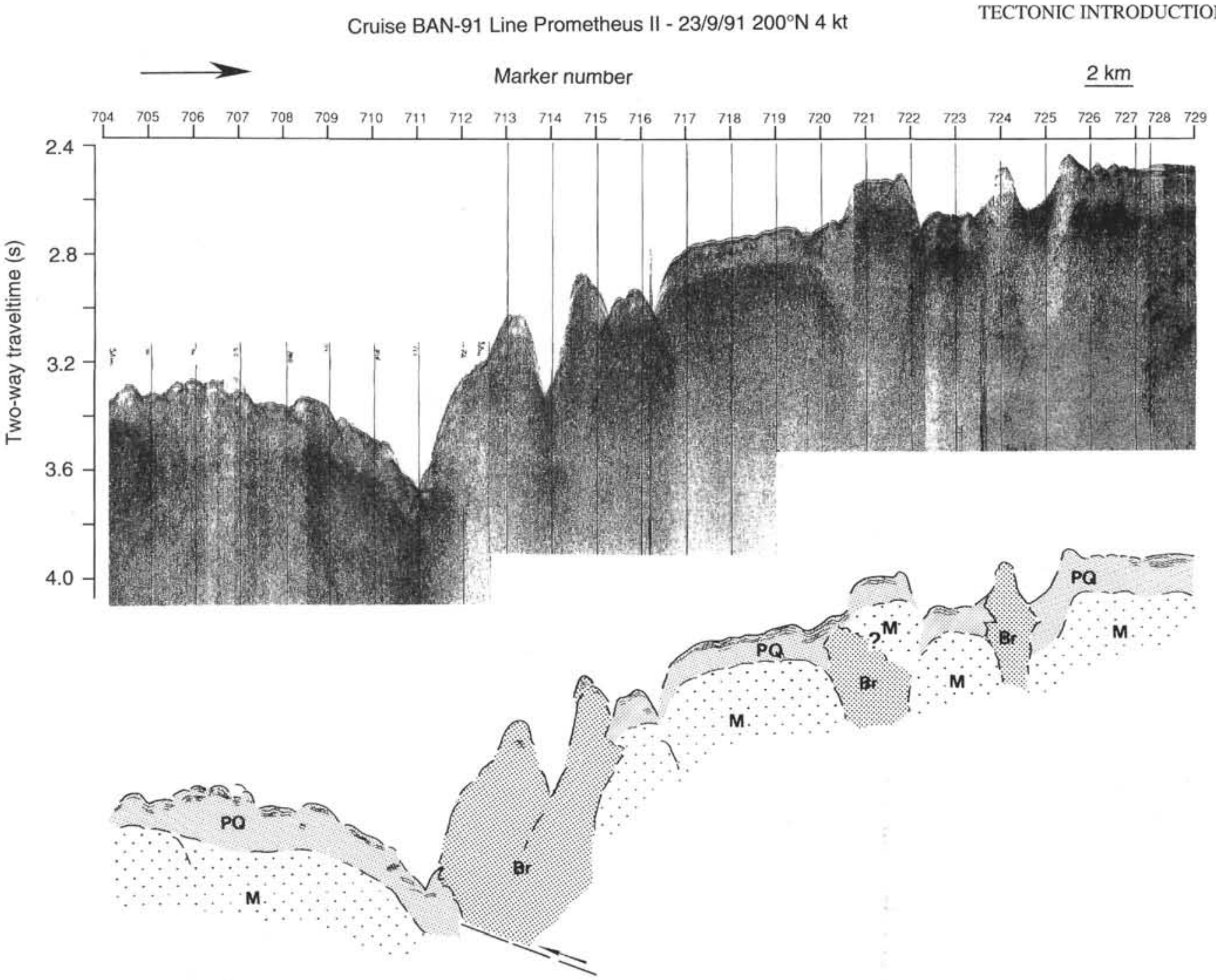

Figure 14. Setting of the Prometheus II mud diapir field. The interpretation of the seismic line in the lower part of the figure shows the inferred diapirs (Br) cutting across the interpreted Miocene and overlying Pliocene-Pleistocene units. Note that the diapirs are also inferred to form above a zone of backthrusting of the Mediterranean Ridge along the Inner Deformation Front. After Camerlenghi et al. (1995).

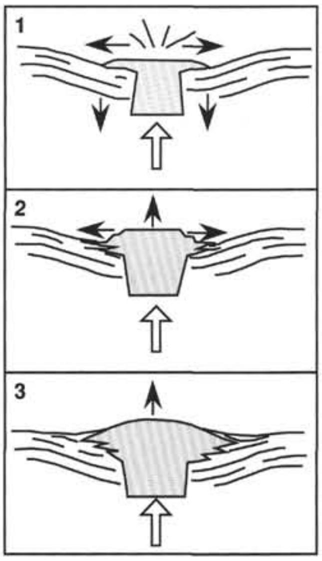

Figure 15. Evolution of mud diapirism leading to the formation of mud volcanoes on the Mediterranean Ridge. The evolution is inferred largely from the interpretation of single-channel seismic data and cores. According to Camerlenghi et al. (1995), the following sequence of events takes place: (1) gas-rich mud is protruded onto the seafloor, which subsides to form a depression; (2) the depression is then infilled with mud extrusions that interfinger with background sediments; and (3) viscous mud is protruded and causes uplift to form a dome-shaped structure. In an alternative model (in Limonov et al., 1994) the dome (stage 3) is formed by upbuilding of mud debris flows, following extrusion from vents in the crestal area, in a manner analogous to the formation of subaerial (magmatic) volcanoes. One of the objectives of drilling at Sites 970 and 971 was to test these two hypotheses.

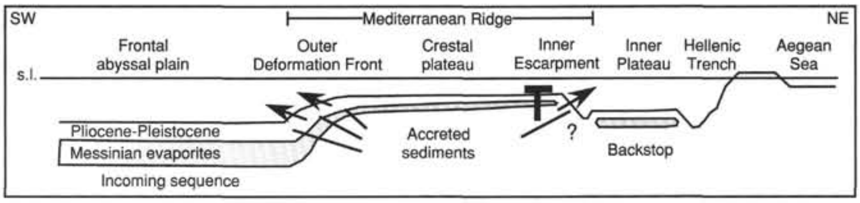

Figure 16. Inferred tectonic setting of mud diapirism on the Mediterranean Ridge. The mud volcanoes are found on the northern edge of the inner crestal plateau along the Inner Deformation Front (Inner Escarpment). The Inner Plateau area is thought to comprise older continental basement rocks that act as a backstop. In this area, fluid pressure builds up at depth, giving rise to mud diapirism and mud volcanism (Camerlenghi et al., 1995).