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Origin of the Eastern Mediterranean basin: a reevaluation

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Abstract

The origin of the Eastern Mediterranean basin (EMB) by rifting along its passive margins is reevaluated. Evidence from these margins shows that this basin formed before the Middle Jurassic; where the older history is known, formation by Triassic or even Permian rifting is indicated. Off Sicily, a deep Permian basin is recorded. In Mesozoic times, Adria was located next to the EMB and moved laterally along their common boundary, but there is no clear record of rifting or significant convergence. Farther east, the Tauride block, a fragment of Africa–Arabia, separated from this continent in the Triassic. After that the Tauride block and Adria were separate units that drifted independently. The EMB originated before Pangaea disintegrated. Two scenarios are thus possible. If the configuration of Pangaea remained the same throughout its life span until the opening of the central Atlantic Ocean (configuration A), then much of the EMB is best explained as a result of separation of Adria from Africa in the Permian, but this basin was modified by later rifting. The Levant margin formed when the Tauride block was detached, but space limitations require this block to have also extended farther east. Alternatively, the original configuration (A2) of Pangaea may have changed by ~500 km of left-lateral slip along the Africa–North America boundary. This implies that Adria was not located next to Africa, and most of the EMB formed by separation of the Tauride block from Africa. Adria was placed next to the EMB during the transition from the Pangaea A2 to the Pangaea A configuration in the Triassic. Both scenarios raise some problems, but these are more severe for the first one. Better constraints on the history of Pangaea are thus required to decipher the formation of the Eastern Mediterranean basin.

Keywords: Eastern Mediterranean basin; Origin; Permian and Early Mesozoic rifting; Adria; Tauride block

1. Introduction

The Eastern Mediterranean basin (EMB, Fig. 1) is a relic of the Mesozoic Neotethys Ocean (e.g., Robertson and Dixon, 1984; Şengör et al., 1984; Dercourt et al., 1986; Le Pichon et al., 1988; Stampfli et al., 2001). To the east and south, its original passive margins are preserved, whereas its present northern and western margins were shaped by later subduction and plate convergence. Seismic refraction studies show that the EMB has an up to 10 km thick probably oceanic crust (and/or strongly attenuated continental crust) overlain by 6 to >12 km of sediment (Makris et al., 1983, 1986; DeVoogd et al., 1992; Ben-Avraham et al., 2002). This is supported by the positive Bouguer gravity anomaly over this basin (IOC,

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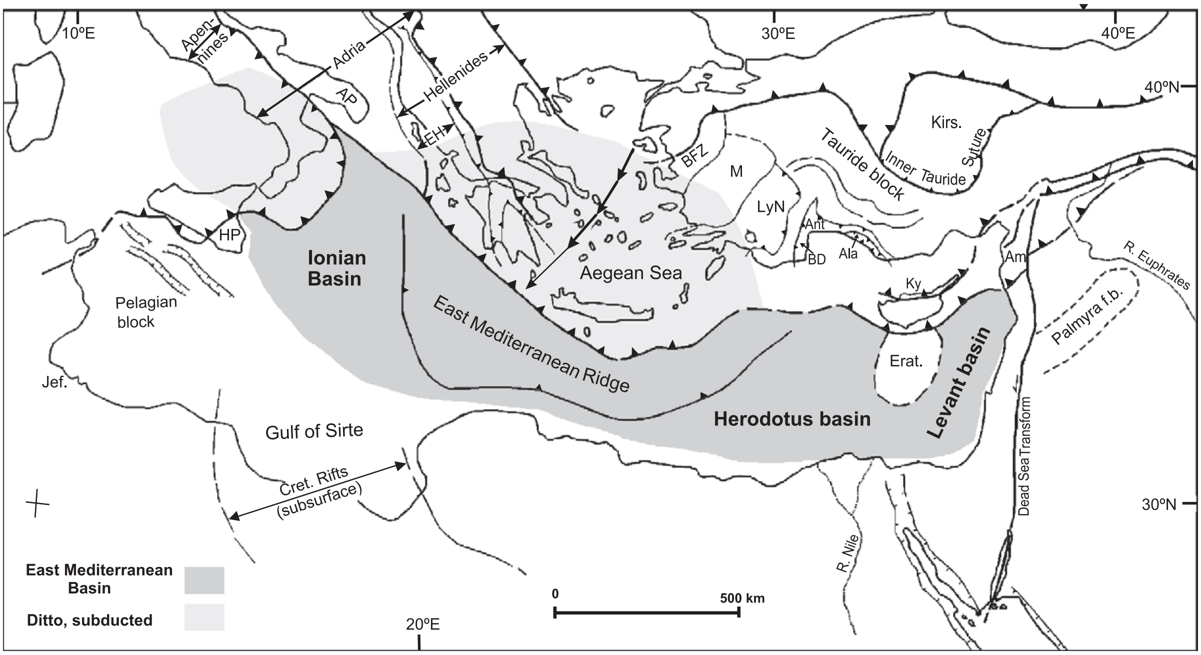

Fig. 1. The East Mediterranean Basin (EMB) and the extent of its subducted parts (only those parts for which there is direct evidence are shown: see text for discussion). Arc with arrows in the Aegean Sea shows the amount of rotation of the margin of the External Hellenides relative to Eurasia. Abbreviations denote Ala., Alanya Massif; Am., Amanos Mountains; Ant, Antalya Complex; AP, Apulian platform; BD, Bey Dağları; BFZ, Bornova flysch zone; EH, External Hellenides; Erat., Eratosthenes block; Jef, Jefara; Kırş., Kırşehir massif; Ky, Kyrenia Range; LyN, Lycian nappes; M, Menderes massif.

1989). In contrast, the Africa–Arabia continent next to the passive margins of this basin has 30- to 35-km-thick continental crust (Ginzburg and Ben-Avraham, 1987; Makris et al., 1988). Such a change in crustal structure allows the interpretation that the EMB formed as a result of rifting, which led to detachment and northward drifting of blocks away from these passive margins. This view is widely accepted, but the history of rifting, the identity and original location of the detached blocks, and the growth history of the basin remain incompletely understood (e.g., Robertson et al., 1996).

The purpose of this work is to reevaluate the origin of the EMB by reexamining and integrating the constraints provided by its passive margins and by the tectonic relations on its northern side. Such an indirect approach is required because there is no direct information about the basin floor. This study expands and updates a previous study (Garfunkel, 1998) by dealing with the entire EMB and stressing the relations between this basin and the blocks on its northern side.

2. Passive margins of the Eastern Mediterranean basin

Important information regarding the origin of the EMB comes from its passive margins. Much of this material has been summarized before (Garfunkel, 1998), so only a short review is presented here.

2.1. The North Arabian margin

The Mesozoic history of the northern Arabian margin is recorded by sediments and volcanics that have been scraped off the sea floor, disrupted and incorporated into melanges and wildflysch, and were then thrust over the edge of the Arabian platform, together with ophiolite nappes, in Late Cretaceous times. In SE Turkey, these rocks show that a continental slope and a deep basin had existed along the northern edge of the Arabian platform since the Middle or Late Triassic, and that then basalt with MORB affinity formed in this basin (Fontaine et al., 1989; Fourcade et al., 1991; Yılmaz, 1993). The north

Arabian margin thus probably began to form no later than ~235 Ma (Anisian). However, rifting may have begun earlier, perhaps at 260–255 Ma, like in Oman (Blendinger et al., 1990; Rabu, 1993). In fact, the northern edge of the Arabian platform began to subside towards this margin in the Permian (Fontaine et al., 1989; Yılmaz, 1993), which supports the existence of a continental margin since then.

In the west, the allochthonous sediments and volcanics and the associated ophiolites extend into NW Syria and still farther west to Cyprus, the continuity of the ophiolite belt under the intervening sea being recorded by short-wavelength magnetic anomalies (Woodside, 1977). In NW Syria, the oldest slope and basinal sediments are Late Triassic and are associated with intraplate basic volcanism, indicating that the Late Triassic continental margin extended to this area (Delaune-Mayere, 1984; Al-Riyami et al., 2000). The Mamonia Complex of Cyprus is similar, but contains Late Triassic MORB as well as intraplate volcanics (Malpas et al., 1992; Robertson and Xenophontos, 1993), indicating the existence of a Triassic oceanic basin. The presence of Early Cretaceous quartzose sandstones in these two basinal complexes suggests an original position not far from the northern Arabian platform. In the northern Arabian platform itself, an Early Cretaceous episode of exposure and erosion followed by sandstone deposition interrupted a long history of carbonate deposition (Ponikarov et al., 1969). In particular, sandstones of Aptian–Albian age (base not exposed) were documented in the NW corner of Syria (Kurd Dag), not far from the ophiolites. Moreover, oil exploration in the nearby part of Turkey revealed exposure of Paleozoic clastics (shales and sandstones) by Early Cretaceous erosion (Temple and Perry, 1962). One of these eroding areas may have supplied sandstones to the adjacent basin in the Cretaceous. Other nearby areas, such as the Taurides in southern Turkey and the Kyrenia Range in northern Cyprus, could not have supplied this sand, because in these localities carbonate deposition continued uninterrupted from Jurassic to Late Cretaceous times (Gutnic et al., 1979; Özgül, 1984; Robertson, 2000a).

Taken together, these lines of evidence suggest that by the Late Triassic a continental margin had already developed along the northern edge of the Arabian platform, the adjacent basin extended to the northern part of the Levant basin, and MORB had already formed in this basin. The Antalya complex of southern Turkey may represent a more northerly part of the same basin (see below).

2.2. The Levant margin

The history of the southern part of the Levant margin off Israel and NE Sinai is well known from boreholes and seismic reflection data (Fig. 2; Druckman, 1984; Garfunkel and Derin, 1984; Druckman et al., 1995; Garfunkel, 1988, 1998). Here, the Arabian platform began to subside towards the present continental margin in the Late Permian. Until the end of the Mesozoic, 6–8 km of shallow water sediment accumulated under the present coastal plain, but only 1 km accumulated some 200 km inland. In contrast, the vertical motions in this region before the Permian were unrelated to the present margin (Garfunkel, 1998), which supports the notion that at this stage this margin did not yet exist.

Significant syn-sedimentary normal faulting took place in mid-Anisian to Norian times and in Liassic to Bajocian times, with a 10–15 Ma long period of reduced tectonic activity and some erosion in between. Where drilling has reached Permian beds (southward of latitude 31°30′ N), minor faulting of that age is compatible with the data, but more significant Permian faulting probably occurred farther north (see below). The total fault offsets were largest, up to 2 km or more, in a belt extending 50–70 km inland of the present coast. The faults are normal and trend subparallel to the coast, so their activity must have resulted in extension at a large angle to the Levant margin. This argues against previous suggestions (e.g., Dercourt et al., 1986; Stampfli et al., 1991) that this margin originated as a N–S trending dextral strike-slip fault. The faulting at this margin was accompanied by igneous activity (Garfunkel, 1989), notably by a >3.5 km thick basalt sequence, mostly 193–197 Ma old, which filled a structural depression south of Haifa (Kohn et al., 1993); older volcanics of Norian age may also be present.

Continental slope and deep-water sediments, younger than the Middle Jurassic, are documented along the southern Levant margin, this being the oldest level reached by nearshore and offshore drilling (Derin, 1974; Kashai et al., 1987; Picard and Hirsch,

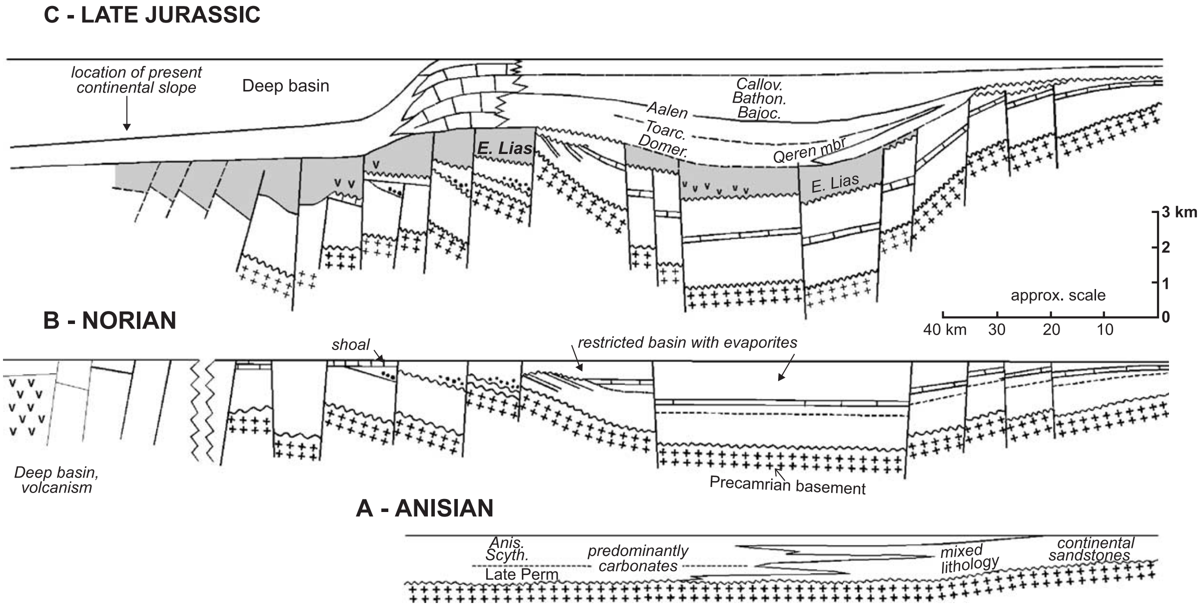

Fig. 2. Schematic history of the south Levant margin until the end of the Jurassic, based on Garfunkel (1998), perpendicular to the southern Levant margin. (A) Situation during the Anisian: no resolvable deformation. Minor Permian faulting may have occurred, but faulting of such age was probably more clearly developed north of this section; (B) situation during the Norian after block faulting. (C) The effects of the Liassic faulting, which accentuated the older block faulting, followed by the Middle and Late Jurassic development of a shallow-water carbonate platform over the faulted substratum next to a deep-water basin.

1987; Druckman et al., 1994). These facies formed basinward of a wide shallow-water carbonate platform whose edge reached near to the present coast (Fig. 2). While in the Middle and Late Jurassic shallow-water carbonates accumulated on the platform, the slope and basin sediments are rich in poorly fossiliferous shales interbedded with calciclastics that were transported downslope (e.g., in a well ~20 km offshore; Druckman et al., 1994). At the end of the Jurassic, the basinward slope along the edge of the shallow-water platform was ≥2 km high, indicating that the adjacent Levant basin was already a deep water basin (Garfunkel, 1998). This argues against suggestions that the Levant basin formed only in the Cretaceous (e.g., Dercourt et al., 1986; Laws and Wilson, 1997) and supports the notion that it is much older (Bein and Gvirtzman, 1977).

The strongly faulted belt next to the southern Levant margin was linked to the intracontinental Palmyra rift in Syria (Garfunkel, 1998; Fig. 1). This rift first formed in mid(?)-Permian times and subsided again in the Triassic, the total subsidence reaching ~2.5 km (Al-Youssef and Ayed, 1992; Beydoun and Habib, 1995; Sawaf et al., 2001). In the west, it joined a deep depression that extended from Galilee to the south of Jerusalem. These structures no longer join, as they are now separated by the ~105 km left lateral offset along the Dead Sea transform. The Triassic section in this depression is >1.7 km thick, showing that it subsided much more than the better studied area farther south (Garfunkel, 1998). Older beds were not drilled, but the link with the Palmyra rift suggests that this depression too was first shaped by Permian faulting.

This history is interpreted as showing that the Levant margin was shaped by Triassic, Early Jurassic, and perhaps also Permian rifting events. The relative importance of these events still cannot be constrained, but a Triassic age for opening of the Levant basin is likely (Garfunkel, 1998). This continental slope was already well shaped in the Middle Jurassic.

There are no data regarding the Middle Jurassic and older history of the northern Levant margin between latitude 32°30′ N and the area in NW Syria considered above. However, like farther south, the adjacent continental area subsided towards the Levant basin from the Middle Jurassic at least (Ponikarov et al., 1969; Sawaf et al., 2001). Moreover, according to Ben-Avraham et al. (2002), the thin (probably, oceanic) crust extends northward along the Levant margin. Therefore, it is assumed that both the northern and southern parts of the Levant margin had similar histories.

2.3. The Northeast African margin

The northeast African margin differs from the south Levant margin in that there is no evidence for Permian to Early Jurassic faulting and basinward subsidence on the adjacent land. In Egypt west of the Nile, Late Liassic beds generally lie directly on pre-Permian beds (although Early Permian beds occur locally), but farther west in Libya, thin Triassic beds are present (Hantar, 1990; Fawzy and Dahi, 1992; Ghori, 1991). Early Jurassic faulting can be identified near the Nile, but has not been documented farther west. Subsidence towards the Mediterranean margin is well expressed only since the Late Liassic. There is little published information on the offshore area. However, an offshore borehole near longitude 23°45′ E (NE Libya) reached deep-water beds of Callovian age, which contain slumped material derived from a nearby shallow-water platform, proving the existence of a continental slope since that time at least (Duronio et al., 1991). A similar situation was revealed by boreholes near the NW Sinai coast (Jenkins, 1990). As there is no structural discontinuity between these areas, it is inferred that the entire passive margin between Sinai and NE Libya was established before the end of the Middle Jurassic and was the continuation of the Levant margin. It is thus inferred that this segment of the EMB margin and adjacent Herodotus basin are at least of that age but may well be older, similar to the Levant basin. The land next to this basin also subsided considerably in the Early Cretaceous and was faulted in this period (Hantar, 1990; Fawzy and Dahi, 1992; Ghori, 1991). However, given the Jurassic subsidence towards the margin and the presence of deep water facies mentioned above, this faulting is interpreted as signifying intraplate deformation, though it could have also affected the continental margin and modified the adjacent basin.

Still farther west, the North African passive margin of the EMB continues WNW across the Sirte Gulf (Fig. 1). Del Ben and Finetti (1991) show a gradual thinning of the African continental crust from the Libyan coast over a distance of >250 km to the Ionian Basin, which is underlain by oceanic crust that they consider to be of Middle Jurassic age. Here the margin of the EMB was considerably overprinted by the NW–SE-trending Cretaceous Sirte rift system (Fig. 1).

2.4. The Central Mediterranean margin

In the Central Mediterranean, the EMB is bordered by the autochthon of Sicily (the Hyblean Platform) and the Pelagian shelf (Fig. 1), which are underlain by continental crust, but this crust thins considerably west of the Hyblean Plateau and the Malta Escarpment (Nicolich et al., 2000). The Pelagian shelf is crossed by a system of prominent Late Cenozoic rifts, under which the crust was thinned to 10–15 km, and it was also affected by Cretaceous faulting accompanied by volcanism, related to the Sirte rifts (Fig. 1; Jongsma et al., 1985; Del Ben and Finetti, 1991). However, older faulting and volcanism are also well recorded in this area and on the adjacent land.

In the Jefara region near the NW Libyan coast, a >4 km thick sequence of marine Permian sediments was drilled (Fig. 1; Lys, 1988). This records the formation of a deep depression, interpreted as a rift. It subsided mainly in Murghabian times (~258–253 Ma ago), but this rift became inactive in the latest Permian. This was an intracontinental structure, being developed inland of the continental crust of the Pelagian shelf offshore. It is unclear whether it extended to the margin of the EMB.

In the Triassic, especially since the Middle Triassic, a wide area west of the Sirte Gulf subsided significantly and was also affected by widespread faulting and volcanism. This activity is known across Tunisia (Bouaziz et al., 1998; Soussi et al., 2000; Bedir et al., 2001; Kurtz, 1983) and in the nearby part of Algeria (Ait Salem et al., 1998). Middle Triassic volcanism of intraplate affinity is also known from the northern Pelagian shelf (Longaretti and Roccho, 1990). Seismic refraction data from northern Tunisia and the adjacent Pelagian shelf (Buness et al., 1992) reveal ~12 km of sediment, mostly pre-Jurassic, which overlies a ~20 km thick crystalline crust near the coast. As the Neogene rifting hardly affected this area, it is possible that the crust was thinned by Early Mesozoic (and/or Permian?) rifting.

In Sicily, the Triassic and Jurassic sections in the autochthon and thrust sheets record syn-sedimentary faulting, sometimes accompanied by volcanism, which produced high-standing platforms and basins (Catalano and D'Argenio, 1978; Pattaca et al., 1979; Longaretti and Roccho, 1990; Catalano et al., 1995). Paleomagnetic data show that the allochthonous units rotated clockwise by up to ~100° during emplacement in the Late Cenozoic (Channel et al., 1990; Oldow et al., 1990; Speranza et al., 2003). Therefore, the palinspastic position of these units is uncertain, but before the breakup of Pangaea they must have been located near the African margin west of the Hyblean Plateau. This emphasizes the extent of the Early Mesozoic faulting. The deposition since the Liassic of deep-water sediments this area suggests accentuated subsidence, probably next to a deep basin (Pattaca et al., 1979; Catalano et al., 1995).

In the present context, a significant feature of this area is the presence of deep-water sediments in the Sicanian allochthonous unit (Catalano et al., 1991, 1992; Di Stefano et al., 1996). The oldest beds in this unit are Middle Permian (~260 Ma) to Late Triassic deep-water and base-of-slope siliciclastics, which contain fragments derived from a nearby shallow-water platform. These are overlain by a basinal series consisting mainly of cherty and nodular carbonates of Late Triassic to Cenozoic age, which indicate the longevity of this deep-water basin. The Permian beds contain a circum-Pacific pelagic fauna, which implies an unrestricted connection to the global ocean. The long stratigraphic record and the persistence of deep-water conditions are compatible with deposition of the Sicanian sequence on strongly thinned continental crust or oceanic crust. Taken together, all these features are interpreted as indicating that this Sicanian basin was part of a broad seaway near North Africa and was directly connected to the Tethys Ocean farther east. Moreover, given its position and longevity, the Sicanian basin is best interpreted as having been continuous with the former western part of the EMB that has subducted beneath Calabria (Fig. 1).

2.5. Summary of passive margins of the Eastern Mediterranean basin

The preceding review indicates that the passive margin of the EMB already existed in the Middle Jurassic and that the adjacent lands subsequently subsided towards the EMB. This is taken as evidence that the EMB dates from that time at least. However, along the Levant margin and in the central Mediterranean region, the available data reveal Permian, Triassic, and Early Jurassic faulting and igneous activity, which record earlier stages of shaping of the EMB margin, suggesting that at least parts of the EMB date from Triassic or even Permian times. However, the land next to the Levant margin subsided considerably towards the margin since the Permian, whereas the coastal region west of the Nile subsided towards the EMB only since the Late Liassic. Such different pre-Middle Jurassic histories raise the possibility that different parts of the EMB margins evolved differently.

3. The northern boundary of the Eastern Mediterranean basin

To better understand the origin of the EMB, information from its passive margins should be supplemented with information about the identity, original location, and drift of the blocks that moved away from these margins and are now located north of this basin. However, the original relations of this basin with the area north of it were obliterated by the continuing tectonic activity along its northern boundary. This raises the question as to whether the present situation reflects the tectonic elements that existed when the basin formed. This summary indicates that the present structural segmentation of the northern side of the EMB probably arose only in mid-Cenozoic times when new subduction zones developed along this boundary, and overprinted and probably also obscured older structures. Therefore, the active structures are not a reliable guide for deciphering older tectonic relations.

At present, the northern boundary of the EMB comprises several distinct segments. The Cyprus Arc, the easternmost segment, is seismically active (e.g., Ambraseys and Adams, 1993) and has been deformed in the Late Cenozoic (Ben-Avraham et al., 1988; Kempler and Garfunkel, 1994; Robertson, 2000a,b). The continuing sedimentation and lack of deformation in southern Cyprus and the young age of uplift of Cyprus show that this segment has been active only since sometime in the Miocene, while earlier in the Cenozoic it was not active (Robertson, 2000a). The boundary of the EMB west of Cyprus does not display much young deformation (Anastakis and Kelling, 1991).

Farther west, the subduction of the EMB below the Aegean (Hellenic) Arc is evidenced by a clear Benioff zone and by tomography (Fig. 1; Richer and Strobach, 1978; Papazachos et al., 2000; Spalman et al., 1988). Consumption of the EMB along the Aegean Arc mainly resulted from the southward retreat of the subducted slab, which led to coeval expansion of the Aegean Sea and clockwise rotation of the External Hellenides (Le Pichon and Angelier, 1979; Kissel and Laj, 1988; Speranza et al., 1995; Fig. 1). This rotation, by 45–50°, began ~25–30 Ma ago, which shows that convergence between the southern Hellenides and the EMB has occurred at least since then, but as there is no clear evidence for older subduction along the Aegean Arc (see below), the convergence along the Aegean arc is constrained to have begun in mid-Cenozoic times.

West of longitude 20°E, the seismically active belt continues into the interior of the Adria microplate along the external Hellenides, whereas the northern boundary of the EMB continues along the foot of the essentially aseismic Apulian Platform (e.g., Kiratzi and Papazachos, 1995; Fig. 1) that is underlain by slightly deformed Mesozoic and Cenozoic sediments (Monopolis and Bruneton, 1982; Charier et al., 1988; Gambini and Tozzi, 1996; Catalano et al., 2001), though a part of it rotated ~20° clockwise in post-Eocene times (Tozzi et al., 1988). Here there is no record of old or ongoing subduction of the EMB beneath the Apulian Platform, but it seems to have been shaped by lateral slip (see below).

Still farther west, the EMB subducts beneath the Calabrian Arc. The ~500 km long Benioff zone under the Tyrrhenian Sea (Fig. 1; Anderson and Jackson, 1987; Selvaggi and Chiarabba, 1995) indicates that formerly the EMB extended westward, forming a corridor between North Africa and the western part of the Adria microplate, and it was consumed in the Late Cenozoic as a result of slab retreat and coeval ESE migration of the Calabrian Arc (Malinverno and Ryan, 1986; Doglioni et al., 1999; Catalano et al., 2001, Fig. 1).

4. Relations with blocks north of the EMB

The preceding considerations show that the present structure of the northern side of the EMB arose only in mid-Cenozoic times when new subduction zones developed. Therefore, to decipher the original relations on the northern side of the EMB, the Early Cenozoic and Mesozoic setting must be examined. During these periods, the area north of the EMB was divided into a western domain that was dominated by the Adria microplate (microcontinent) and the Anatolian domain farther east. For the present discussion, two issues are important: the nature of the pre-Cenozoic boundary between the EMB and these domains, and the relations of the blocks north of the EMB with its passive margins.

4.1. The Adria microplate

This microplate (also called Apulia) has behaved as a coherent unit since the beginning of the Mesozoic at least. Regarding its pre-Jurassic position, two points are relevant. First, the presence of Variscan basement in the northern Apennines and the Southern Alps (Vai and Cocozza, 1986) shows that Adria was attached to the Variscan orogen of Europe, as is widely accepted. Moreover, the presence of Variscan basement in Crete in the Phyllite–Quartzite nappe, whose low structural position indicates derivation from the margin of Adria (Seidel et al., 1982; Finger et al., 2002), shows that at least a large part of Adria has Variscan basement. Second, the Phyllite–Quartzite nappe of Crete contains clastics with open-marine pelagic Permian and Triassic fossils (Krahl et al., 1983, 1986). Catalano et al. (1991, 1995) recognized their similarity to the coeval sediments of the Sicanian basin and concluded that they were deposited on the northern side of this seaway. This implies that the southern margin of Adria already existed in the Permian and Triassic, when a deep seaway separated it from Africa.

Several lines of evidence bear on the subsequent (i.e., post-Early Jurassic) positioning of Adria relative to the EMB. First, evidence from the passive margins of the EMB, presented above, shows that this basin separated Adria and Africa since that time at least, which supplements the evidence regarding the Sicanian seaway. Second, the Early Jurassic position of Adria is constrained by the reconstruction of Pangea obtained by closing the Atlantic Ocean. Though several variants have been proposed (e.g., Van der Voo, 1993), they all require placing Adria ~>500 km east of its present position relative to Africa to avoid overlap of continental areas. This implies that sometime after the Early Jurassic Adria moved a considerable distance westward with respect to Africa and the EMB (which was attached to Africa) (as shown, for example, by Dercourt et al., 1986). Third, paleomagnetic data do not reveal resolvable changes in the position of Adria relative to Africa since the Permian, from which it has been inferred that Adria has remained rigidly attached to Africa (Channel, 1996). This is incompatible, however, with the reconstruction of Pangea that indicates some motion between them, as noted above. Nonetheless, these constraints can be reconciled by assuming a left-lateral displacement (~500 km) of Adria along its boundary with the EMB, which would not be resolved by paleomagnetic data.

Such a displacement is compatible with the lack of recognizable evidence along the southern margin of Adria for Mesozoic motion perpendicular to the boundary between this margin and the EMB. There is no record of Jurassic or later rifting and passive margin formation, such as normal faulting or basinward thickening of the Mesozoic sequence, as would be expected according to some models that assume such rifting (e.g., Dercourt et al., 1986). On the other hand, the abrupt transition over a distance of ~10 km (Catalano et al., 2001) between the Apulian platform and the EMB, discussed above, is consistent with the interpretation that it has been shaped by the lateral motion during the westward shift of Adria, which was discussed above. Such motion would modify a preexisting (pre-Jurassic) passive margin along the southern boundary of Adria. Neither is there evidence for subsequent subduction along this margin. Though tectonic erosion accompanying such subduction (Von Huene and Scholl, 1991) could obliterate the evidence, the lack of deformation (e.g., folding), the absence of igneous activity, and the paleomagnetic evidence argue against significant Adria–EMB convergence in the Mesozoic.

Another relevant feature is the internal deformation of Adria. In the Triassic, it was fractured (rifted) and differentiated into several basins separated by relatively high-standing platforms. In the west, the Lagonegro basin, interpreted as a rift, formed in the area from which the Apennine nappes were later derived (Wood, 1981; Iannace and Zamparelli, 2002). Northwest-trending (relative to present orientations) Triassic rifting also occurred along the axis of the Adriatic Sea (Grandić et al., 2002), while farther east the Ionian basin and its flanking platforms formed in the southern Adriatic area and in Greece (Jacobshagen, 1986; Robertson et al., 1991; De Alteriis and Aielo, 1993; Zappaterra, 1994). These structures are not related to the southern margin of Adria, as they extend into the interior of Adria and thus trend at an angle to this margin. The considerable Mesozoic subsidence of Adria may be related to this rifting, but no relation between this subsidence to the southern margin of Adria has been recognized. The Triassic rifting that formed the Pindos Ocean along the eastern margin of Adria (Jones and Robertson, 1994; Dengan and Robertson, 1998; Pe-Piper, 1998) also followed the trend of these intra-Adria structures.

Since the Late Eocene, strong compression sliced the eastern and western parts of Adria into thrust sheets, which were displaced towards Adria's slightly deformed central part, forming the Apennines and the External Hellenides. These mountain chains trend oblique to the Adria–EMB boundary (Fig. 1). Thus, their formation cannot be related to convergence along the Adria–EMB boundary, which supports the absence of subduction along this boundary until then. Subduction along the eastern part of this boundary became important only since the Aegean arc developed in mid-Cenozoic times.

Taken together, these lines of evidence show that since the Permian the southern margin of Adria was separated from Africa by a deep seaway that developed into the EMB. A few hundred kilometers of lateral motion along the Adria–EMB boundary may have occurred after the Early Jurassic, but there is no evidence for either significant extension or plate convergence across this boundary in the Mesozoic. The geometry of internal deformation of Adria is also unrelated to subduction along this boundary.

4.2. The Anatolian domain

In contrast to Adria, the Anatolian domain comprised several distinct microcontinents that were separated by Neotethyan oceans. North–south plate convergence since the Late Cretaceous has led to collision of these microcontinents and has eliminated the intervening oceanic areas, vestiges of which are represented by ophiolitic sutures (e.g., Şengör and Yılmaz, 1981; Dilek et al., 1999). In the present context, the major questions regarding this domain are: what were the original locations of these microcontinents; and how did they drift?

The Tauride (or Anatolide–Tauride) block is the largest in the Anatolian domain (Fig. 1). It is accepted as a fragment of Gondwana because it is underlain by Paleozoic sediments, mainly clastics, which resemble the coeval sediments of northern Arabia and North Africa (e.g., Ricou et al., 1975; Gutnic et al., 1979; Demirtaşlı, 1984; Özgül, 1984). The absence of Paleozoic deformation in this block, and the Late Precambrian Pan-African basement that is widely exposed in the Menderes Massif (e.g., Bozkurt and Oberhansli, 2001), and Late Ordovician glacial deposits (Monod et al., 2003) are also characteristic of northern Gondwana. Moreover, the long-lasting supply of clastics shows that this block was a part of a large continental area that could provide this material. The Early Paleozoic sequence of the Tauride block resembles that of the northern Arabian platform, so this allows an initial position either next to the Levant margin, as is often assumed, or next to more eastern parts of the Arabian margin. However, the presence of glacial deposits (Monod et al., 2003) favors the first possibility, as proposed by Garfunkel (1998), because farther east glacial deposits extend only as far as northern Jordan.

The Antalya complex (Fig. 1) provides evidence of Triassic rifting and the formation of a deep basin along the southern side of the central Tauride block, which record its separation from Gondwana (Robertson et al., 1991; Robertson, 2000b). The presence of Late Triassic lavas transitional between intraplate basalt and MORB in the Antalya complex suggests initiation of an oceanic basin south of the Tauride block at that time (Robertson and Waldron, 1990). This basin probably connected with the coeval oceanic basin recorded in Cyprus. However, the Late Permian fauna of the Tauride block records a restricted environment on its southern side (Altiner et al., 2000), which suggests that then an open seaway did not exist there. On the other hand, there is considerable evidence for deformation of the Tauride block coevally with the rifting events recorded by the Antalya Complex.

Though incomplete exposure of Triassic sediments, and their typical severe tectonization into a pile of thrust sheets, do not allow a complete palinspastic reconstruction of the Tauride block in that period, the detailed descriptions of many areas (Gutnic et al., 1979; Demirtaşlı, 1984; Demirtaşlı et al., 1984; Özgül, 1984; Monod and Akay, 1984) provide important indications. The outstanding feature is that the Middle and Late Triassic sediments of the Tauride block show great variability in thickness and facies in different structural units such as adjacent thrust sheets. These sediments comprise shallow water carbonate-dominated sections, sandstone series, and conglomerates, while thicknesses vary by hundreds of meters. In places igneous activity occurred. The sandstone occurrences, which are limited in areal extent, indicate localized uplift, exposure, and erosion of sources of the clastics (most likely Paleozoic sediments). Sporadic occurrences of clasts of Permian limestone record the creation and erosion of local relief. Taken together, these features record considerable syn-deposition tectonically controlled differential vertical motions, which may reach hundreds of meters. These led to uplift and erosion in some areas and to differential subsidence in other areas, which controlled the thickness variations of the sediment. This picture is best interpreted in terms of fault-controlled block tectonics. In addition, Monod and Akay (1984) and Özgül (1984) show that different parts of the Triassic section were unconformably covered by Jurassic sediments after having been eroded, which further records the Triassic structural differentiation of this region. They also bring evidence from a few places for compressional deformation in the latest Triassic or earliest Liassic. This deformation appears to be post-depositional, as it affects sizeable sections, and thus post-dates the syn-depositional tectonism discussed above. From the end of the Triassic on a deep basin already existed south of the Tauride block, as evidenced by the Antalya Complex, so the deformation of this block during this period could have been decoupled from that of the area to the south.

South of the Tauride block, several continental splinters are present. One is the Alanya block, which has been thrust northward over the Antalya complex

(Fig. 1; Okay and Özgül, 1984; Robertson, 2000b). Thus, a part of the basinal area from which the Antalya complex was derived originally intervened between the Alanya and Tauride blocks. The Kyrenia range, northern Cyprus, has a distinct history (Baroz, 1979; Robertson, 2000a), so it may also be a part of a separate unit, but continuity with the Tauride block cannot be ruled out. Makris et al. (1983) and Ben-Avraham et al. (2002) inferred from seismic refraction data that the Eratosthenes block south of Cyprus (Fig. 1) is a continental splinter with thinned crust, up to 20 km thick, with ~5 km thick cover. A prominent magnetic anomaly over this block indicates that its cover contains voluminous volcanics (probably of Early Mesozoic age; Garfunkel, 1998). Zverev and Iliinskii (2000) inferred a different seismic velocity distribution and suggested that this block is essentially a volcanic edifice. The original positions of these small blocks are only constrained as south of the Tauride block.

To constrain the history of drifting of the Tauride block, its mid-Cretaceous position relative to the EMB will be investigated, based on the closure history of the Neotethyan seaways in Anatolia (e.g., Yılmaz, 1993; Yiğitbaş and Yılmaz, 1996; Polat et al., 1996; Parlak and Delaloya, 1999; Dilek et al., 1999; Okay and Tüysüz, 1999; Robertson, 2000b). Most of these Neotethyan seaways began to close in Cenomanian–Turonian time (95–90 Ma)—the ages of metamorphic soles that formed as a result of intra-basinal thrusting, when convergence between Africa and Eurasia first began (e.g., Dewey et al., 1973; Le Pichon et al., 1988; Müller and Roest, 1992). Later in the Cretaceous, when subduction and basin closure were more advanced, melange formation and fragmentation of ophiolite bodies became widespread, but final ophiolite obduction and basin closure typically occurred only in the Tertiary. This history allows the tectonic relations some 90 million years ago to be constrained (Fig. 3). Although different arrangements of the subduction zones are also possible, this will not change the main conclusions.

The emplacement of the Troodos—NW Syria (Baer–Bassit)—SE Turkey (Hatay) ophiolite belt and the elimination of the basin in which they formed record a north-dipping subduction zone that consumed part of the EMB and its eastward continuation north of the Arabian platform (e.g., Robertson and Xenophontos, 1993; Yılmaz, 1993; Yiğitbaş and Yılmaz, 1996; Dilek et al., 1999; Robertson, 2000b). It was active for ~20 million years, until the Early Maastrichtian (~70 Ma), and was eliminated when the ophiolites were emplaced onto the north Arabian margin and over what remained of the Levant basin. The higher allochthonous units in SE Turkey contain Late Cretaceous and younger deep-sea sediments and ophiolites (Yılmaz, 1993; Yiğitbaş and Yılmaz, 1996; Dilek et al., 1999; Robertson, 2000b), which indicate that farther north an additional oceanic area—here called the south Tauride basin—existed between this subduction zone and the Tauride block (Fig. 3). Thus, while obduction of the Troodos–Hatay ophiolite belt records elimination of the basin and subduction zone south of this belt, another basin and subduction zone persisted farther north into the Eocene (Yılmaz, 1993; Yiğitbaş and Yılmaz, 1996). A Senonian magmatic arc along the northern side of this basin indicates that then a north-dipping subduction was already active along the northern side of this basin. Persistence of the basin into the Eocene shows that it was consumed over a period of ~40 Ma, so most probably it was rather wide. Convergence between the Tauride block with Arabia has continued into the Late Cenozoic.

West of where the Dead Sea transform meets the belt of Alpine deformation, the record of the south Tauride basin is less clear, but here too there is evidence for basin closure. Yiğitbaş and Yılmaz (1996) record emplacement in the Eocene of ophiolites over the edge of the Arabian platform in the Amanos Mountains (Fig. 1). Farther west, Miocene sediments in the Misis Range (NE continuation of Kyrenia block) contain debris derived from an island arc (Floyd et al., 1992). Their position is compatible with derivation from the westward continuation of the south Tauride basin. The Late Eocene thrusting of the Kyrenia Range over Troodos is also interpreted as recording the closure of an intervening basin, because the very different histories of these units show that originally they were far apart (Baroz, 1979; Robertson, 2000a). Farther west, basin closure led to north-directed thrusting of the Antalya Complex onto the southern margin of the Tauride block in the Early Eocene, but the basin from which this complex originated began to close in the Cretaceous (Robertson et al., 1991; Robertson, 2000b). The Alanya Massif, thrust over the Antalya Complex (Fig. 1),

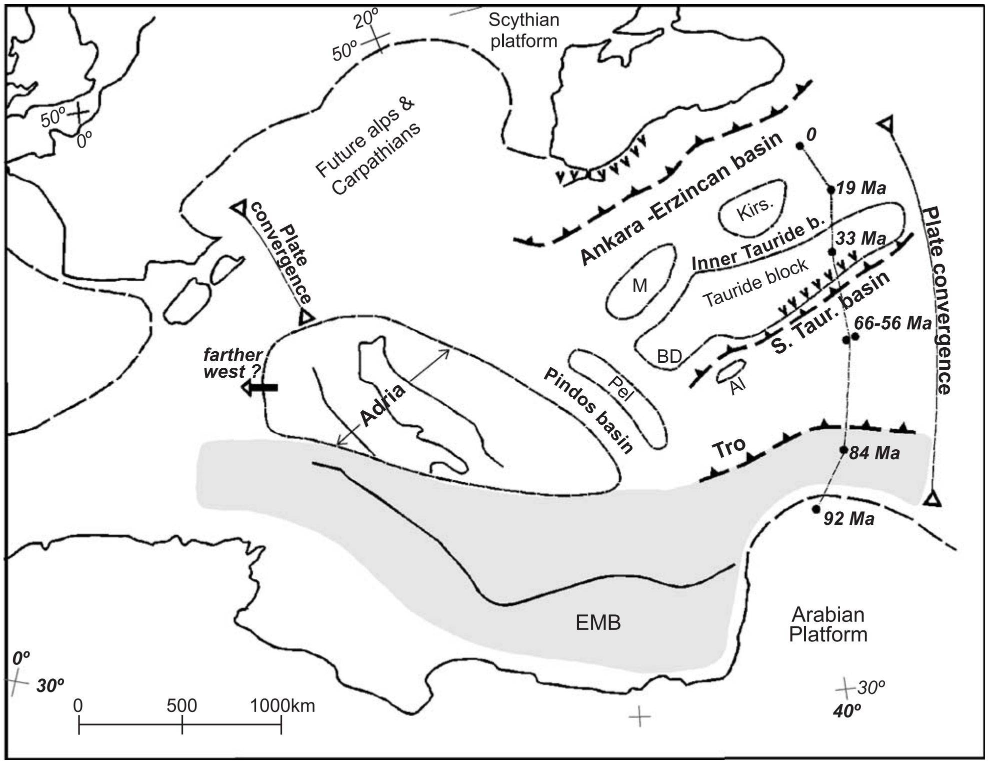

Fig. 3. Sketch showing the tectonic relations north of the EMB some 90 Ma ago, before closure of the Neotethyan seaways within the Anatolian domain. Note the interpreted continuity of structures as far west as the Menderes Massif/Bey Dağları (and adjacent localities that are now within islands in the SE Aegean Sea), then a major discontinuity farther west before the Pelagonian Block is reached. The Lycian Nappes may have originated from the western continuation shown of the Inner Tauride basin, between the Menderes Massif and Bey Dağları, or from the Ankara–Erzincan basin farther north, the latter interpretation being preferred by many authors (e.g., Robertson, 2000b). Relative positions of Africa and Eurasia are according to Müller and Roest (1992). Abbreviations are the same as in Fig. 1. See text for discussion.

experienced high-pressure blueschist metamorphism (Okay and Özgül, 1984), suggesting significant plate convergence (i.e., subduction or major continental overthrusting). This indicates the presence of a suture south of the Tauride block, like farther east.

Yet another oceanic basin north of the Tauride block—the inner Tauride basin—was probably the source of the ophiolites at the top of the Hoyran–Hadim nappes (S and SW of the Inner Tauride Suture; Fig. 1) that were thrust over the northern margin of the central Tauride block in the Late Eocene (Gutnic et al., 1979; Özgül, 1984; Polat et al., 1996; Dilek et al., 1999; Andrew and Robertson, 2002). Farther west, the Lycian nappes (Fig. 1; Gutnic et al., 1979; Collins and Robertson, 1998; Robertson, 2000b) were thrust on the Bey Dağları sector of the Taurides in the Miocene. However, they were previously tectonized in the Late Cretaceous and record the closure history of an oceanic basin along a northwest-facing continental margin (relative to present-day orientations). The basin was interpreted to have been located north of the Menderes Massif (Gutnic et al., 1979; Şengör and Yılmaz, 1981; Collins and Robertson, 1998, Robertson, 2000b), NE of this massif (e.g., Okay and Tüysüz, 1999; Okay et al., 2001), whereas Özkaya (1990) suggested that the ophiolites originated north of the massif while the underlying carbonates south of it. However, an origin of all nappers from a seaway between the Menderes Massif and Bey Dağları (Fig. 3) cannot be excluded. In the present context, the important thing is that in all these interpretations the Lycian nappes originated from a westward continuation of one of the basins of the Antolian domain discussed above, and their history records roughly N–S Cretaceous–Palaeogene plate convergence at the longitude of the Menderes Massif (Fig. 3).

Still further north was the Ankara–Erzincan branch of the Neotethys, whose closure produced a very prominent suture (Figs. 1 and 3; e.g., Şengör and Yılmaz, 1981). This seaway was consumed by a north-dipping subduction zone that was probably already active in the Aptian (Okay et al., 1996). North of it, a calc-alkaline magmatic arc developed in mid-Turonian to end Campanian times, and final collision along this suture zone occurred in the Early Tertiary (Fig. 3; Okay and Tüysüz, 1999; Okay et al., 2001).

This brief summary shows that the plate convergence across the Anatolian domain since the mid-Cretaceous has been partitioned between several subduction and suture zones. The relative motions between the African and Eurasian plates (e.g., Müller and Roest, 1992) require ~1800 km of plate convergence across this domain since 90 Ma, ~800 km of which post-dates the Early Eocene (~56 Ma) (Fig. 3). It seems likely that the convergence south of the Tauride block was not less than north of it. This view is supported by the existence of two subduction zones south of it and the magmatic arc along its eastern part, and because the duration of plate convergence south of this block was not less than north of it. A conservative estimate is thus that at least one-third of the plate convergence took place south of the Tauride block, such that at ~90 Ma this block was located at least 600 km north of its present position relative to the EMB (Fig. 3).

4.3. The relation between Adria and the Tauride block

Many models of the Neotethyan history (e.g., Robertson and Dixon, 1984; Dercourt et al., 1986; Stampfli et al., 1991, 2001) assume that Adria was attached to the Tauride block to form a single elongated microcontinent, so these units moved in unison. This view is based on the concept that the zones of similar facies (isopic zones in the older literature) of the external Hellenides extend all the way across the southern Aegean Sea, although the correlation is not straightforward (e.g., Aubouin et al., 1976b; Bonneau, 1984). On the other hand, Bernoulli et al. (1974) doubted the validity of such a correlation, and Hall et al. (1984) proposed that the zones of similar facies of the Hellenides do not extend east of Crete. The previously estimated mid-Cretaceous position of the Tauride block (Fig. 3) also does not fit the single microcontinent concept. To examine this issue, which is important for the history of the EMB, the major aspects of the Late Cretaceous to Early Tertiary histories of the two sides of the southern Aegean region are compared, although a full regional analysis is beyond the scope of this work. Fig. 4 thus shows the pre-Neogene relations between the different units in this region, with the effects of the young extension in the Aegean region removed.

In the west, the Hellenide nappes extend from Peloponnese to Crete (Fig. 4; e.g., Jacobshagen, 1986; Bonneau, 1984; Papanikolaou, 1988; Seidel et al., 1981). These include the external units derived from Adria, the Pindos nappe derived from an oceanic deep-water basin east of Adria (the Pindos basin in Fig. 3), and the allochthonous Late Jurassic (Eohellenic) ophiolite bodies that were originally emplaced on the eastern side of this basin (Smith, 1993; Jones and Robertson, 1994). Like elsewhere along the Hellenides, here too flysch deposition in the Pindos Basin began near the end of the Cretaceous and continued until the Late Eocene when this basin closed and its fill was thrust southwestward to form the Pindos nappe. Flysch deposition then shifted to the marginal part of Adria and continued until the Early Miocene when this margin was broken into several thrust sheets that were thrust westward to form the External Hellenides (the Tripolitza and Ionian nappes) (Jacobshagen, 1986).

The islands east of Crete (Fig. 4) record a different history. The highest units in Karpathos (Xinodothio Nappe) and in Rhodes (Profitis Ilias Nappe) include basinal Mesozoic sediments similar to those of the Pindos nappe. However, unlike the actual Pindos nappe, they are associated with melange and fragmented ophiolites of Late Cretaceous rather than Late Jurassic age (Aubouin et al., 1976a; Hatzipnagitou, 1988; Koepke et al., 1985), while Eohellenic ophiolites are not observed. The presence of Late Cretaceous ophiolites, unknown in the Hellenides, links these islands to the Anatolian domain, to the Lycian nappes in particular. Moreover, on Karpathos, a shallow marine Middle Eocene molasse overlies these rocks as well as the underlying Mesozoic limestone

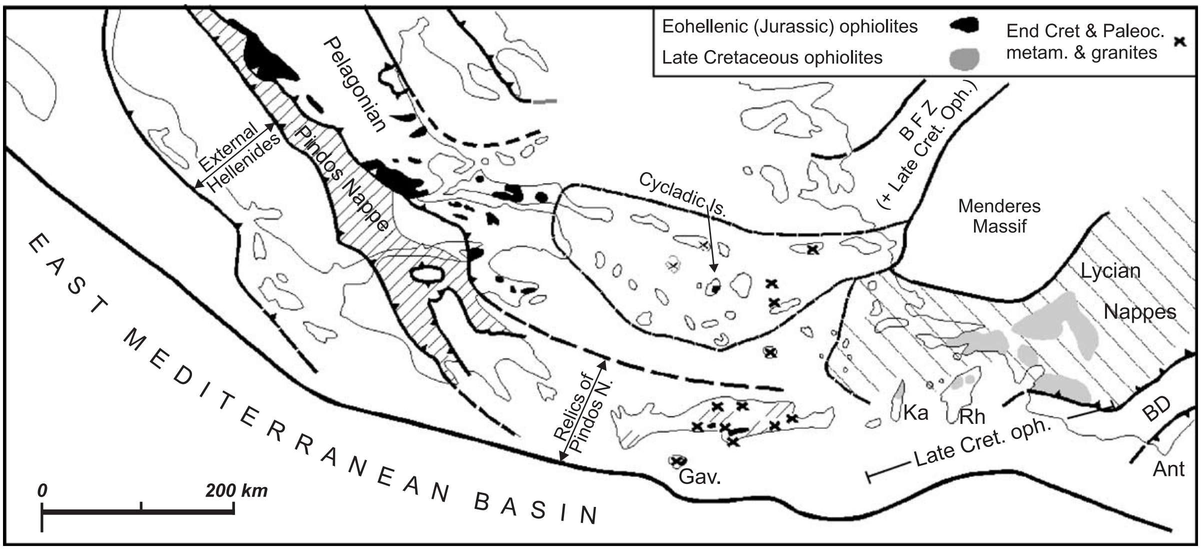

Fig. 4. Major elements on the two sides of the southern Aegean Sea. Young extension of Aegean Sea and rotations of blocks were corrected following Le Pichon and Angelier (1979) and Kissel and Laj (1988), so this diagram provides an indication of the structure around the start of the Neogene. Abbreviations denote Ant., Antalya complex; BD, Bey Dağları; BFZ, Bornova flysch zone; Cr, Crete; Gav, Gavdos; Ka, Karpathos; Rh, Rhodes. In this interpretation, in the Mid-Cretaceous, the crust now forming the Pelagonian and Pindos zones was widely separated from the crustal blocks now in SW Turkey, but these regions became juxtaposed by Late Cretaceous and Early Tertiary motions. The Cyclades Islands are interpreted as underlain by another distinct continental fragment, of unclear provenance, which has Variscan age crust. Remnants of this continental fragment, which is presumed to have become caught up in the relative motions of the surrounding blocks, are also found on Crete.

unit (Kalilimni unit), showing that these units were already stacked while flysch deposition was still going on in the Pindos basin farther west. Thrusting of Middle Eocene age took place, however, in the Lycian Nappes, where broadly coeval sediments overlie previously tectonized thrust sheets (Collins and Robertson, 1998).

These Pindos-like basinal units overlie two carbonate-rich units, the lower Adra unit and overlying Kalilimni unit on Karpathos, and the lower Attavisros unit and overlying Archangelos unit on Rhodes (Mutti et al., 1970; Aubouin et al., 1976a). Like in the External Hellenides (i.e., in the Tripolitza and Ionian Nappes), these units comprise carbonate sections that persist into the Eocene and are overlain by Middle and Late Eocene to Oligocene flysch. However, Aubouin et al. (1976a,b) noted that these carbonate sections differ in facies from those of the External Hellenides. Moreover, nappe stacking on Karpathos was completed before deposition of the Middle Eocene molasse, while in the External Hellenides it began only towards the end of the Eocene (Jacobshagen, 1986), casting further doubt on the continuity of these units with the External Hellenides farther west. More likely is continuity with the lower Lycian nappes farther east, which also record carbonate deposition that continued into the Eocene (Gutnic et al., 1979; Collins and Robertson, 1998). Bernoulli et al. (1974) indeed identified the typical sections of some Lycian Nappes in the islands in the SE Aegean west of Rhodes (Fig. 4), which further supports the affinity of this area with the Anatolian domain.

A rock assemblage peculiar to the central–southern Aegean comprises Maastrichtian to very Early Paleocene (78–63 Ma) high-temperature metamorphic rocks, greenschists, and granites (Seidel et al., 1981; Reinecke et al., 1982; Altherr et al., 1994). They occur as small isolated slices on several islands, where they overlie the Cycladic high-pressure metamorphic complex or Eocene flysch, while in Crete they are imbricated with Eohellenic rock assemblages (Fig. 4). They record the former existence of a large exhumed metamorphic terrain in this area, but its structural position is not clear.

In interpreting these differences, it is possible to assume a priori that the External Hellenic zones of similar facies continued uninterrupted eastwards, which requires the character of these zones to change east of Crete (Aubouin et al., 1976a,b). The other option, favored here, is to infer the presence of two domains with different histories. In the eastern domain, the ophiolites are Late Cretaceous, and much convergence took place in Late Cretaceous to Early Tertiary times, but there is no record of an Eohellenic phase. In contrast, in the western domain, the ophiolites are Late Jurassic, and Late Cretaceous tectonism is inconspicuous or absent.

The estimated motions of the main blocks also conflict with the assumption that in the mid-Cretaceous the western part of the Tauride block was aligned with SE Adria. The location of the latter is constrained by the evidence that it did not move much (in the N–S direction) relative to the EMB, whereas the earlier (conservative) estimate of the position of the Tauride block places it considerably farther north (Fig. 3). These considerations imply that Adria, together with the Pindos basin east of it, was kinematically decoupled from the more eastern domain along a line or zone passing east of Crete. This could have been a transform fault, but the presence in the central Aegean region of end-Cretaceous high-grade high-temperature metamorphics (Seidel et al., 1981; Reinecke et al., 1982; Altherr et al., 1994) and of the blueschists in the Cyclades suggests that subduction along this boundary also took place. Ring et al. (1999) have also argued for a major discontinuity in the Middle Aegean, based on the age difference between the Variscan basement of the Cyclades and the Pan-African basement of the Menderes Massif.

5. Discussion

These considerations reveal several important constraints on the origin of the EMB. Most important is the evidence showing that its passive margins already existed in the Middle Jurassic. Moreover, where enough information is available, rifting events in the Permian, Triassic, and Early Jurassic are recorded along these margins. There is also evidence that oceanic basins existed next to the Levant and the northern Arabian platform from the Triassic at least, and north of Sicily since the Permian. These inferences suggest that the EMB originated in the Permian or the Early Mesozoic, when the major continents were still assembled as Pangaea.

5.1. The Adria–EMB relation

The role of Adria in the formation of the EMB is crucial because the original size of Adria, when the shortening during the formation of the Apennines and Hellenides is restored, was comparable to the EMB. Accepting that in the Permian and Early Mesozoic Adria was attached to the Variscan orogen of Europe at the western end of the Tethys Ocean, its position with respect to Africa can be inferred from the reconstruction of Pangaea. Its configuration in the Early Jurassic (Pangaea A) is obtained by closing the Atlantic Ocean (e.g., Van der Voo, 1993). In all variants of this restoration Adria must be placed east of its present position relative to Africa to avoid overlap of continental areas (Fig. 5), as was noted above. This places Adria next to the site of the EMB, so the Sicanian basin on its southern side appears to be the precursor of the EMB.

The Sicanian Basin could originate in two ways. First, it may have formed by Permian rifting followed by drifting of Adria away from Africa. However, since Adria was a part of the Variscan orogen (see above; Fig. 5), this would have caused considerable deformation on its northern side, which has not been identified. Neither has the rifting expected along the southern margin of Adria been identified. Second, the Sicanian Basin may be a relic of a Paleozoic ocean that did not close during the assembly of Pangaea, like the Gulf of Oman that did not close during the Alpine orogeny. This implies that the southern margin of Adria was shaped by Paleozoic subduction (which could have led to tectonic erosion rather than piling up of an accretionary prism; cf. Von Huene and Scholl, 1991). The difficulties with the first interpretation thus no longer arise. On the other hand, this revised interpretation implies that much of the North African margin of the EMB formed in pre-Permian times, making it much older than any other existing continental margin. This cannot be tested on the basis of existing data, but it seems unlikely.

Interpretation of the Adria–EMB relations must, however, also take into account the paleomagnetic

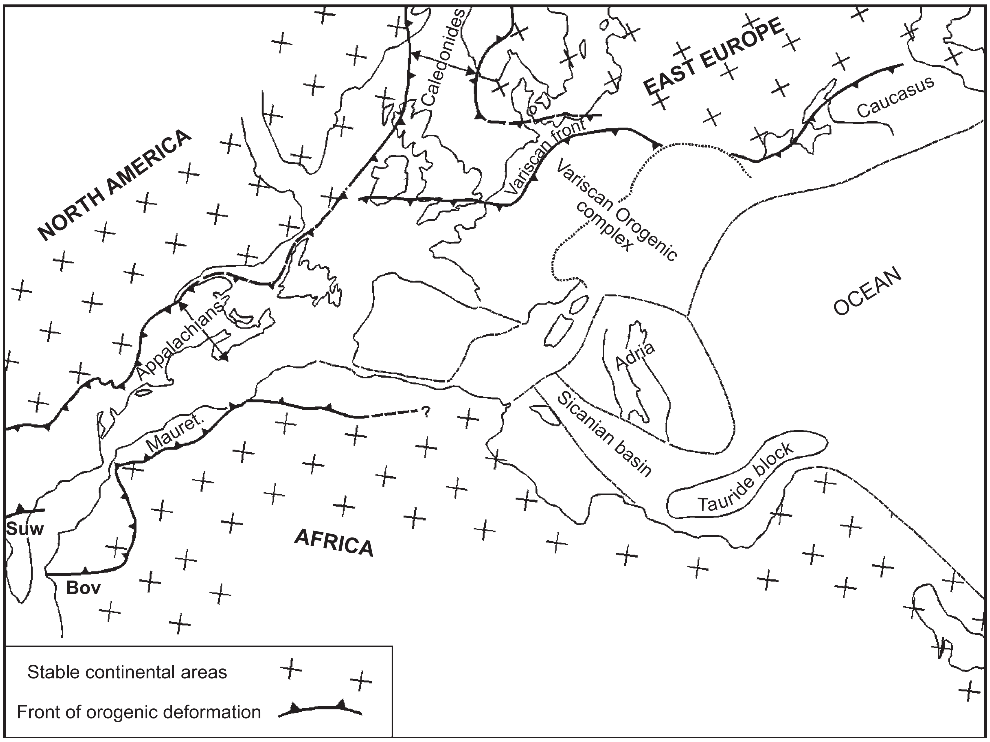

Fig. 5. The palinspastic setting of the Mediterranean region at the end of the Permian, with major continents arranged in the Pangaea A configuration. Abbreviations denote Suw, Suwanne terrane and basin; Bove, Bove basin.

studies that suggest that the configuration of Pangaea changed. Though geological data show that most parts of Pangaea accreted in the Late Carboniferous, Pangaea-A type reconstructions do not match the pre-Jurassic apparent polar wander paths of Laurussia and Gondwana (Smith and Livermore, 1991; Van der Voo, 1993). The relative positions of the major components of Pangaea may thus have changed sometime in the Permian–Triassic, but models invoking large changes (e.g., Pangaea-B configurations) have been considered unlikely because it has not been clear how they can evolve to a Pangaea-A type configuration. Van der Voo and Torsvik (2001) and Torsvik and Van der Voo (2002) showed that this problem could be largely resolved if the geomagnetic field included variable non-dipole components. However, even with this assumption, to fit the paleomagnetic data they invoked a Late Permian reconstruction that differed from Pangea A (Torsvik and Van der Voo, 2002, their Fig. 9). In this configuration (called Pangaea-A2; Smith and Livermore, 1991; Van der Voo, 1993), Gondwana is rotated clockwise with respect to Laurussia relative to the situation at the beginning of the Jurassic. The transition from Pangaea A2 to Pangaea A thus requires a right-lateral motion of ~500 km along the Africa–North America boundary.

Supporting evidence is provided by correlating the southeastern USA with West Africa. It has been recognized (Chowns and Williams, 1983; Williams and Hatcher, 1983; Horton et al., 1989; Thomas et al., 1989) that the Mesozoic–Cenozoic sediments in Florida and southern Georgia are underlain by a terrane derived from Gondwana—the Suwannee terrane. This has a Pan-African basement covered by slightly deformed Early Paleozoic sediments with faunas of Gondwanan, not North American affinities. This terrane did not experience Paleozoic orogenic deformation, but was accreted onto North America only in the Late Paleozoic when North America collided with Gondwana. It is thus separated from the Appalachian orogen by a major suture, which extends eastward into the Atlantic Ocean at latitude 32°N. The continuation of this suture can be recognized in West Africa (Lécorché et al., 1983; Dallmeyer, 1989; Figs. 5 and 6). Here, the Mauritanides form the eastern side of the Late Paleozoic orogenic belt produced by the Laurasia–Gondwana collision. They are separated from the older part of Africa by a prominent suture, which bends westwards into the Atlantic Ocean around latitude 13°N. The area east and south of this suture has a Pan-African basement, covered by slightly deformed Early Paleozoic sediments of the Bove Basin. The fill of this basin was correlated in detail with the Early Paleozoic sequence of the Suwanee Basin in SE USA (Chowns and Williams, 1983; Thomas et al., 1989; Villeneuve, 1988). Both these areas were little affected by the extension during the opening of the Atlantic, so they preserve the Paleozoic configuration.

In the Pangea A reconstruction, there is a mismatch of ≥500 km between these tectonic elements (Fig. 5). This suggests that left-lateral slip by ~500 km subsequently occurred along the North America–West Africa boundary, similar to what was inferred from the paleomagnetic data. Two independent lines of evidence thus support a Pangaea-A2-like initial configuration of Pangaea, which later evolved into the Pangaea A configuration by shearing along this boundary.

This possibility has significant implications for the EMB. It leads to a Permian reconstruction in which Adria, attached to the Variscan orogen of Europe, is located west of its present position relative to Africa (Fig. 6), rather than opposite the passive margin of the EMB (Fig. 5). In this reconstruction, only the eastern part of the southern margin of Adria faces Africa and the Sicanian Basin between them is rather narrow; it could thus be a small relict Paleozoic seaway that connects to the Tethys Ocean farther east. This initial position of Adria does not require its detachment from

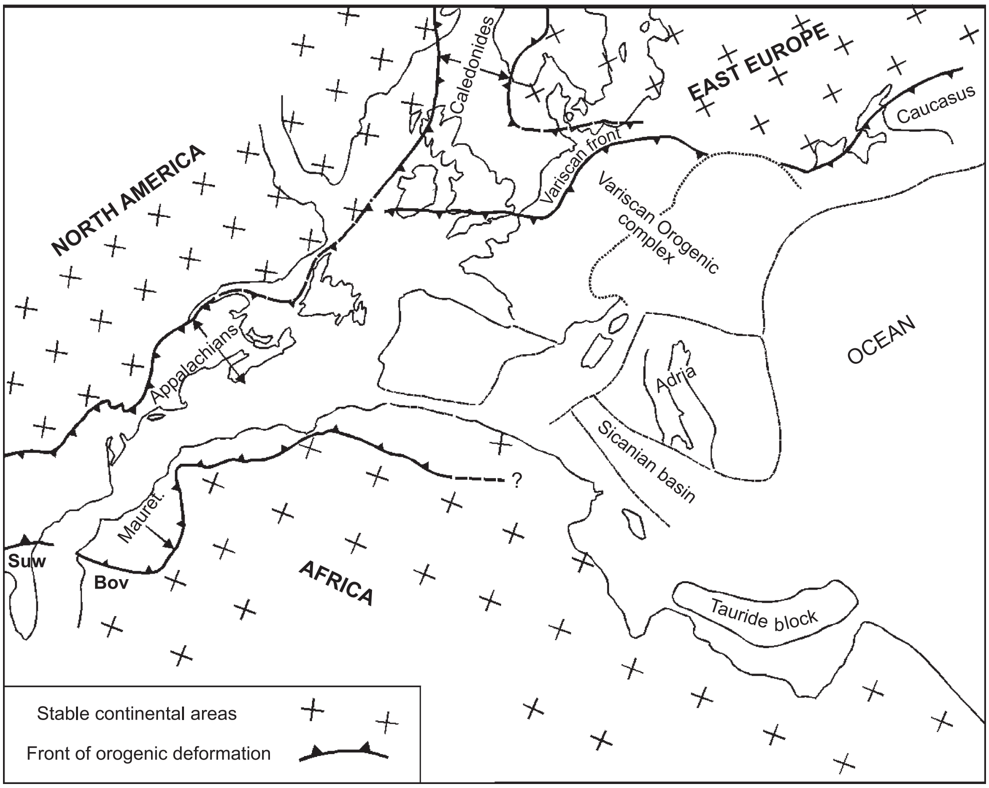

Fig. 6. The position of the East Mediterranean area in a Pangaea A2-like reconstruction of the continents. Note that in this configuration the geologic structures match across the southern Central Atlantic Ocean and that the oceanic area north of the Tauride block is much wider than in Fig. 5. Abbreviations denote Suw, Suwanne terrane and basin; Bove, Bove basin.

Africa to form the EMB, so that the difficulties mentioned above, regarding this detachment, do not arise. However, this interpretation implies that Adria was juxtaposed with the EMB only during the transition from Pangaea A2 to Pangaea A.

5.2. Implications for the origin of the Eastern Mediterranean basin

This discussion suggests that the constraints regarding the history of the EMB, listed above, are still insufficient for formulating a unique model of its origin, especially because its early history is not sufficiently constrained. The basin could have formed in two paleogeographic settings. The first is the Pangaea A setting, with Adria located next to the much of the EMB; the second is the Pangaea A2 setting, with Adria located next to the western extremity of the EMB.

In the first setting, Permian rifting may have separated Adria from Africa to form the Sicanian Basin—the precursor of EMB. Although this model faces some difficulties, as outlined above, it is thought more plausible than the alternative in which the North African margin is pre-Permian. In either case, this basin was modified and most likely widened by Triassic and Jurassic rifting, which is well recorded in the central Mediterranean region. Rifting of these ages was indeed very widespread, having affected the entire Mediterranean region and the boundary between West Africa and North America. In this model, the southern margin of Adria and a large part of the North African margin are conjugated, but their relative positions were modified by the northwestward lateral shift of Adria relative to Africa and the EMB (which occurred close to the Jurassic–Cretaceous transition according to Dercourt et al., 1986). However, there is no good evidence for large motions perpendicular to the EMB–Adria boundary.

Separation of Adria from Africa cannot explain the formation of the entire EMB, because Adria does not seem large enough to extend to the Levant margin (Figs. 5 and 6), so it is inferred that some other block was detached from the eastern part of this basin. This could have been the Eratosthenes block (if indeed it is a continental splinter) or some other minor unit, or it could have been the Tauride block. However, there is not enough room along the passive margins of the EMB to place both Adria (having its original size) and the Tauride block. Thus, if the latter was located next to Adria (with its western part north of the Eratosthenes block), a part of it must have extended along the north Arabian margin (Fig. 5). The arguments presented above showing that Adria and the Tauride block did not form a continuous unit imply that they could have separated independently from the EMB margins, possibly at different times.

In any case, the Triassic and Liassic fault pattern along the southern part of the Levant margin records, as shown above, that the block that separated from this margin moved SE–NW. This sense of motion would be difficult to envisage if Adria and the Tauride block moved in unison, but it could be achieved if they were independent.

In a Pangaea A2-like setting (Fig. 6), the Sicanian Basin between Sicily and Crete can be interpreted as a relict Paleozoic seaway. Farther east, in this configuration, a wide oceanic area intervenes between Adria and the margins of the EMB (Fig. 6). This leaves enough room for placing the Tauride block next to the margins of the EMB. The Tauride block is the only known large unit whose geology is similar enough to that of the lands next to these margins, so it is the most likely candidate to have been originally juxtaposed with the NE African and Levant margins. In this situation, there is no obstacle against the Tauride block moving NW away from the Levant margin without interfering with Adria, before the latter approached the EMB margins when Laurasia and Gondwana were rearranged to produce the Pangaea A configuration in the Late Triassic. It is still possible that separation of small splinters, such as the Eratosthenes block (if it is indeed continental), or the Alanya massif, was involved in the formation of Levant basin, whereas the Tauride block, located further outboard, was rifted separately away from these splinters. The early opening could thus satisfy the kinematic constraint provided by the faults next to the southern Levant margin, and it may also explain the differences in subsidence histories of the Levant and NE African margins.

This model (updating a previous analysis, by Garfunkel, 1998) starting with the Pangaea A2 setting does not encounter the difficulties that arise if one assumes that the Pangaea A configuration existed through the life span of this supercontinent. However, it predicts considerable deformation (maybe even subduction) along the Adria–EMB boundary during the transition from Pangaea A2 to Pangaea A. This idea thus still needs to be tested against better data regarding the Permian–Triassic history of the southern margin of Adria. In addition, the Pangaea A2 configuration and the evidence supporting a large right lateral shear along the original Africa–North America boundary need additional study.

6. Conclusion

This analysis has revealed several constraints bearing on the major questions regarding the origin of the EMB. The age constraints appear to be the best established and require that the EMB formed within the framework of Pangaea before final separation of Gondwana from Laurasia. It is argued that Adria and the Tauride block—the main units adjacent to the EMB—were separated and moved independently during this separation. Between the Jurassic and the Early Cenozoic, significant lateral motions took place along the boundary between Adria and the EMB, even though there is no recognizable evidence for either rifting or convergence across this boundary. The record from the Sicanian Basin, which existed since the Permian, and the Middle Jurassic or older age of the NE African and Levant passive margins of the EMB indicate that the boundary between Adria and the EMB existed in Early Mesozoic and earlier. This contrasts with the record of Triassic rifting on the southern side of the Tauride block and with evidence showing that following its northward drift away from Gondwana considerable subduction and plate convergence occurred along its southern side while it moved 600 km or more towards Arabia.

However, these inferences are insufficient to identify with certainty the original locations of Adria and the Tauride block and how they behaved during the formation of the EMB. For that, a well-constrained reconstruction of Pangaea during formation of the EMB is essential. The reconstruction in which the relative position of Gondwana and Laurasia did not change (Pangaea A) leads to a model in which most of the EMB formed by Permian separation of Adria followed by additional Triassic and Jurassic rifting, but this model involves several difficulties. These difficulties do not arise if the EMB began to form when these major continents were arranged in the Pangaea A2 configuration. In such a model, Adria was originally situated mostly west of the EMB and is brought against it only in the Triassic, while most of the EMB may have formed as a result of the separation of the Tauride block from Gondwana. This highlights the importance of testing the evidence supporting the Pangaea A2 reconstruction and of obtaining more information regarding the southern margin of Adria, especially its Permian–Triassic history. Such improved knowledge will help to decide between different models of the formation of the EMB.

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