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The Mediterranean Ridge backstop and the Hellenic nappes

X. Le Pichon *, S.J. Lallemant [1], N. Chamot-Rooke, D. Lemeur, G. Pascal

Laboratoire de Géologie, UMR 8538, Ecole Normale Superieure, 24 rue Lhomond, 75231 Paris Cedex 05, France

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Abstract

The core of the Mediterranean Ridge backstop consists of a pile of Hellenic nappes that migrated outward from the Aegean continent within the adjacent Mediterranean basins during late middle Miocene, about 15 Ma, and the present Mediterranean Ridge has developed since that time by accretion of a new wedge. The arguments we use are: (1) a similarity in thickness, seismic velocity and structure between the backstop and the Hellenic nappes in southern Peloponnesus; (2) the geometry of the westward limit of the backstop that is the one expected from the amount of relative displacement of the nappes on land; (3) an agreement between the estimated ages of the present accretionary wedge and the period of activity of two major faults connecting the backstop to the adjacent Aegean continent. The seaward limit of the backstop is situated about 170 km southwest of western Crete where the displacement of the nappes is maximal and it is close to the margin near the Ionian islands where their displacement is small. We assume that the Hellenic subduction zone migrated southwestward relative to Eurasia during middle Miocene and that the corresponding gravity collapse of the Aegean continent was maximum at this time because the westward extrusion of Anatolia had not yet started.

Keywords: Mediterranean Ridge; Hellenic nappes; accretion; subduction zones

1. Introduction

The Mediterranean Ridge was discovered by Heezen and Ewing (1963) and interpreted by Emery et al. (1966) as resulting from the convergence between the Mediterranean seafloor and the Hellenic Arc. Rabinowitz and Ryan (1970) proposed that the Ridge is a thickened crustal wedge. Le Pichon et al. (1982) proposed instead that it is a large sedimentary wedge involving the upper 4 km of the sedimentary basins. They pointed out the importance of the geometry of the backstop and assumed that the extension of the lower Hellenic continental margin acts as the backstop. Independently, Ryan et al. (1982) proposed that the Mediterranean Ridge is an accretionary wedge where the decollement level progressively migrates downward from the base of the Messinian evaporites near the front of the wedge to the pelagic Aptian below the top of the wedge. However, very little was known about the nature of the backstop and its location until Truffert et al. (1993) and Lallemant et al. (1994) used wide-angle seismics to establish the existence and geometry of the backstop northeastward of the summit of the Mediterranean Ridge (Figs. 1 and 2).

[1] Present address: Département des Sciences de la Terre, CNRS ESA 7072, Université de Cergy-Pontoise, 8 av du Parc, Le Campus, Bat I, 95031 Cergy Pontoise Cedex, France.

* Corresponding author.

E-mail address: lepichon@geologie.ens.fr (X. Le Pichon).

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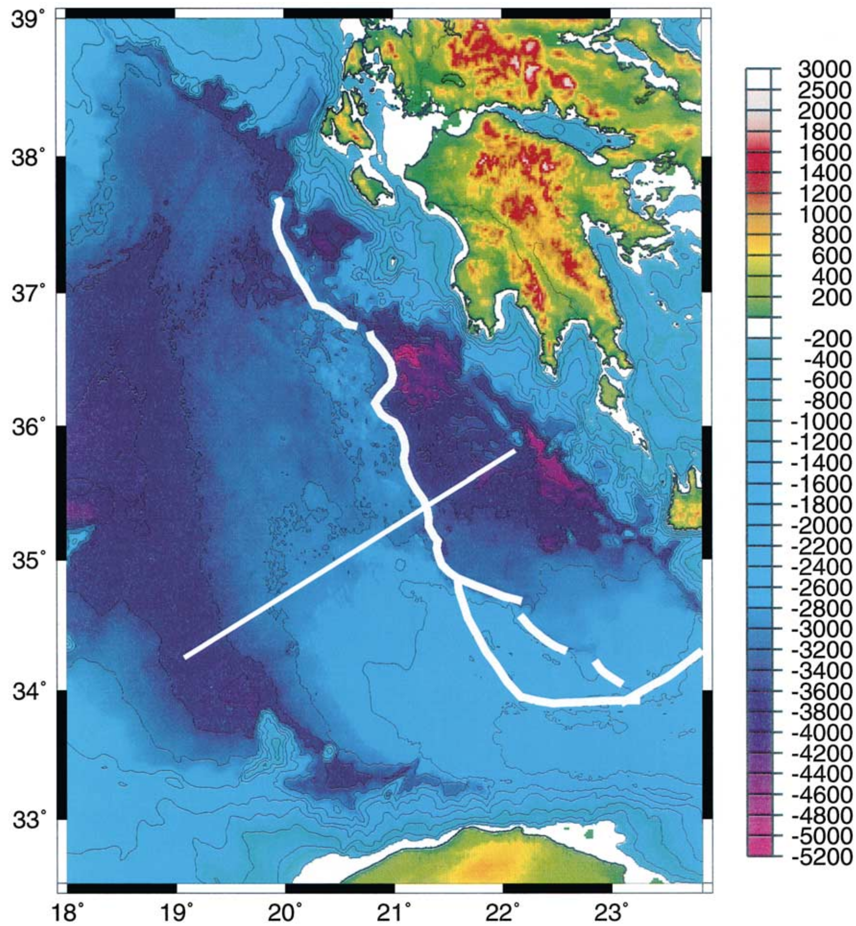

Fig. 1. Topographic map of the Ionian sea area with location of the IMERSE section (Fig. 2B). Bathymetry from Heralis and Médée cruises. The white line is the location of the crustal cross-sections of Fig. 2.

Truffert et al. (1993) and Lallemant et al. (1994) showed that the main part of the backstop consists of a high seismic velocity (4.8–6.6 km/s) material 6.5 km thick sitting on top of lower velocity (4.2 km/s) sediments that they interpreted as being subducted below the backstop (Fig. 2). DSDP 377 borehole within a trough near the backstop leading edge drilled through middle Miocene turbidites (between 19 and 15 Ma) probably of African origin. These turbidites were covered with 15–14-Ma pelagic marls (see discussion in Kastens, 1991). At the time the turbidites were deposited, the present accretionary wedge could not have existed if the turbidites indeed came from Africa. Thus the backstop appears to have been present before 15 Ma and the change of sedimentation to pelagic marls at 15 Ma is compatible with an initiation of the present accretionary wedge at about this time.

Much more information is now available about the configuration and seismic structure of the backstop and the adjacent accretionary wedge including in particular the dense IMERSE seismic network. This information confirms the interpretation of Lallemant et al. (1994) of the backstop as a high velocity slab with a slightly greater thickness (8 km instead of 6 km) (Le Meur et al., 1997; Jones et al., 2002; Reston et al., 2002). The backstop is made of high velocity material (highly indurated sediments or crustal rocks) directly overlain by a thick Messinian evaporite sequence and layered Plio–Quaternary sequences, although we cannot rule out the possibility of a thin layer of pre-Messinian sediments (Jones et al., 2002). In any case, the sedimentary basins overlying the high velocity body appear to be not older than the late Miocene, indicating an emplacement of the backstop in late middle Miocene. This late middle Miocene age is compatible with the volume and structure of the present accretionary wedge (Kastens, 1991; Chaumillon and Mascle, 1997; Reston et al., 2002). We conclude that the backstop reached its present configuration in upper middle Miocene and that the adjacent accretionary wedge was initiated immediately after, near 15 Ma.

An important point is that the backstop keeps a constant thickness of about 8 km over its 100 km width to the foot of the Hellenic margin where the crustal thickness rapidly increases to about 30 km (see Fig. 2). This constant thickness crustal body is thus much thinner than the adjacent continental crust. Yet, there is no indication of major extensional structures that may account for a thinning by a factor of 3 from the Aegean crust to the backstop crust.

Because of the lack of exposure of the high velocity body, we can only speculate on the nature of the backstop. The seismic velocities and thickness are compatible with oceanic crust (although a bit slow). They are also compatible with continental crust or highly indurated sedi-

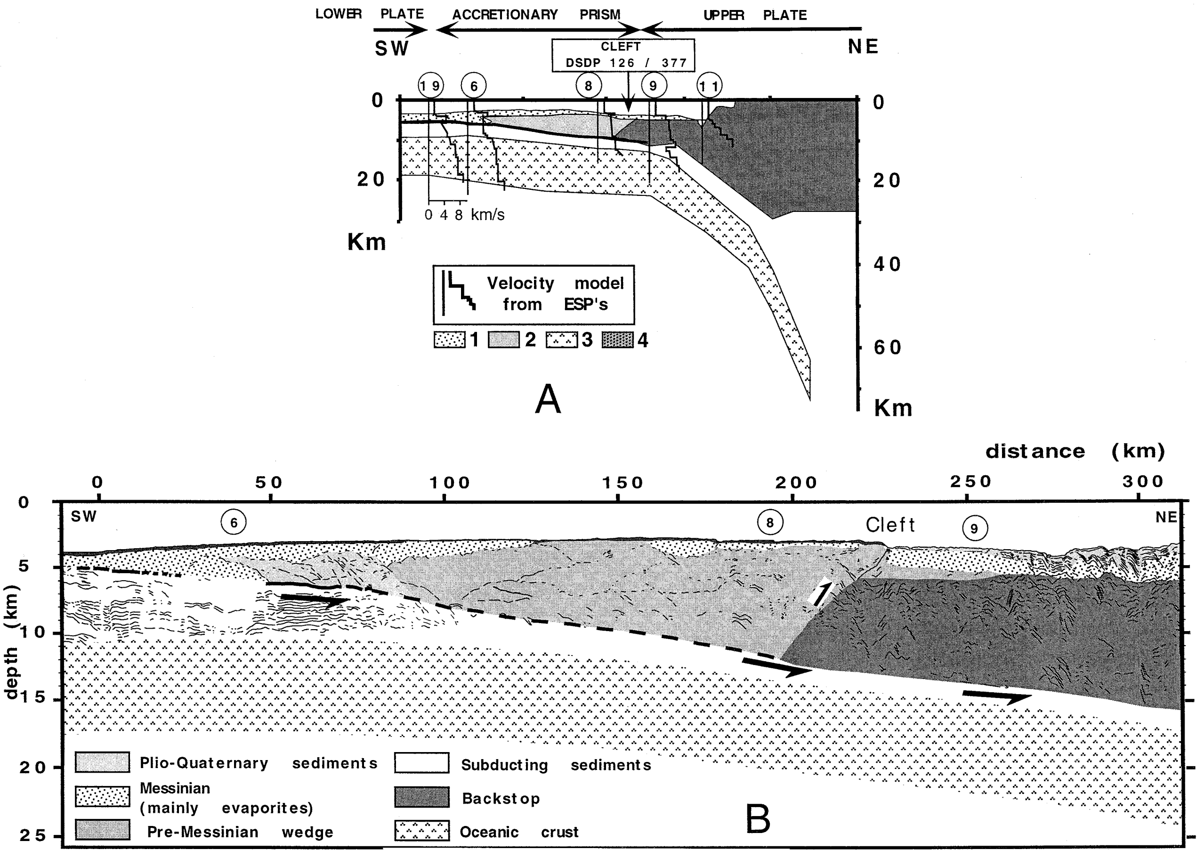

Fig. 2. (A) Crustal cross-section of the Aegean subduction complex (modified from Truffert et al., 1993). Circled numbers are ESP. (B) Synthetic geological section from the IMERSE cruise multichannel section (modified from Reston et al., 2002). Location of the IMERSE profile in Fig. 1, the crustal section of Truffert et al. (1993) being roughly along the same track.

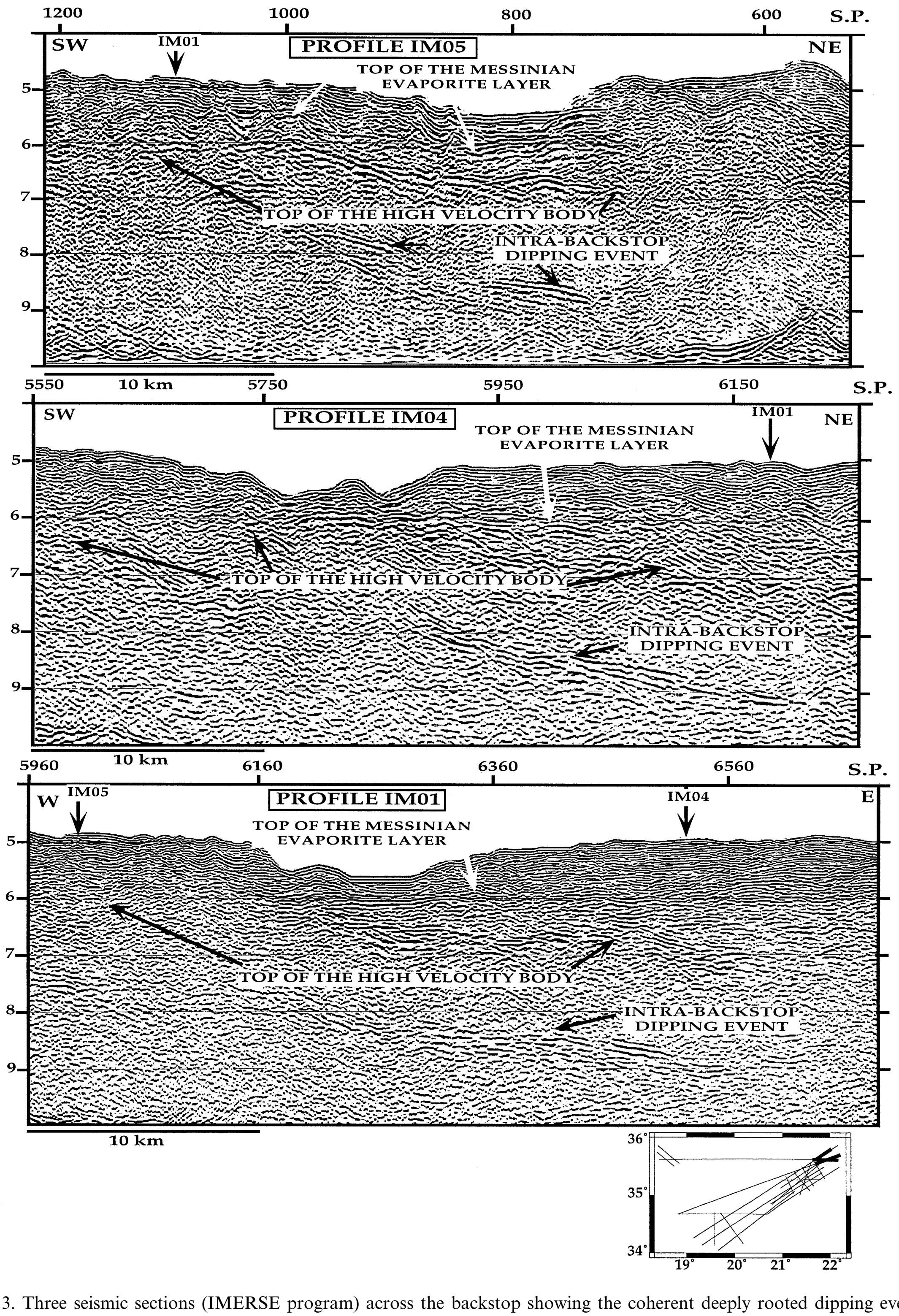

Fig. 3. Three seismic sections (IMERSE program) across the backstop showing the coherent deeply rooted dipping events.

ments of the Hellenic nappes. Low-angle (10–15°) laterally coherent seismic events crossing the whole high velocity body appear to control the distribution of the overlying Messinian basins, although they have no surface expression (Fig. 3). A possible interpretation of these reflectors would be low-angle overthrusts that would favor the Hellenic nappes origin of the basement.

Chamot-Rooke et al. (personal communication, 1996) found that the reflectivity of the backstop obtained with the EM12 multibeam sounder during the Médée cruise is strikingly different from the reflectivity of the sedimentary wedge. They used this difference to map the outer limit of the backstop (see Fig. 1). This rectilinear limit extends about 170 km to the southwest of western Crete but gets progressively closer to the continental margin toward the northwest. At the latitude of the southern Peloponnesus, the distance to the margin is only 70 km and stays relatively constant northward to the latitude of the Kephalonia transform fault. Thus we now have important new sets of independent data on the internal structure, the geometry and the age of emplacement of the very large crustal body that forms the Mediterranean backstop; the aim of our paper is to discuss the implications of this new information.

In a first section, we present the kinematics of the subduction zone using the Sandwell free-air anomaly map derived from Geosat and ERS1 altimetry data and the geodetic data from Greece (Kahle et al., 1995; Le Pichon et al., 1995). In a second section, we show that the outline of the backstop coincides with the expected location of the seaward extension of the Hellenic nappes. We show further that the thickness and seismic velocity of the backstop are compatible with those of the Hellenic nappes. We thus propose that the backstop was formed by migration of a stack of Hellenic nappes from the Aegean continent within the adjacent Mediterranean basin in late middle Miocene time. In a third section, we show that a simple rigid rotation of the nappes forming the backstop can describe the kinematics of this migration. We discuss further two major transverse strike-slip faults extending from the continent to the backstop. These were active during upper middle Miocene time and we show that their geometry is compatible with the simple rotation just described, which suggests that they acted as transform faults at the time the backstop reached its present location. Finally, we discuss the geological significance of this interpretation within the Miocene evolution of the area.

2. The present-day Hellenic subduction: its kinematics and relationship to the Aegean continental margin

South of 37°N, the Sandwell free-air gravity anomaly map in Fig. 4 shows that the backstop coincides approximately with a northwest–southeast rectilinear 450 km long negative gravity trough at the foot of the similarly rectilinear western Aegean continental margin. The gravity trough terminates to the southwest, along the extension of the South Cretan Trough fault system, near 34°N and 24° E (see Fig. 7), and to the northeast, at the latitude of the southern Peloponnesus (near 37°N). The northwestern edge of the negative gravity trough is occupied by a series of topographic troughs, known as the western Hellenic trenches (the deepest being the Matapan Trench). North of 37°N, the backstop coincides roughly with the lower Peloponnesus continental margin and is associated with a positive free-air gravity anomaly instead of the important negative one observed south of 37°N.

The arrows in Fig. 4 show the motion vectors of Aegea with respect to Africa obtained from geodetic measurements after Kahle et al. (1995) and Le Pichon et al. (1995). The motion of the Aegean continental margin is everywhere toward the south-southwest, in general agreement with the focal mechanisms from interplate earthquakes (Taymaz et al., 1990). The velocity decreases by a factor of two from about 40 mm/yr in the south to 23 mm/yr near the Ionian islands. The decrease in velocity occurs discontinuously along the Corinth Gulf and along its extension to the southwest and this discontinuity is related to the north–south opening of the Gulf of Corinth at 15 mm/yr (Le Pichon et al., 1995). Kahle et al. (1995) have shown that north of the Kefallonia trans-

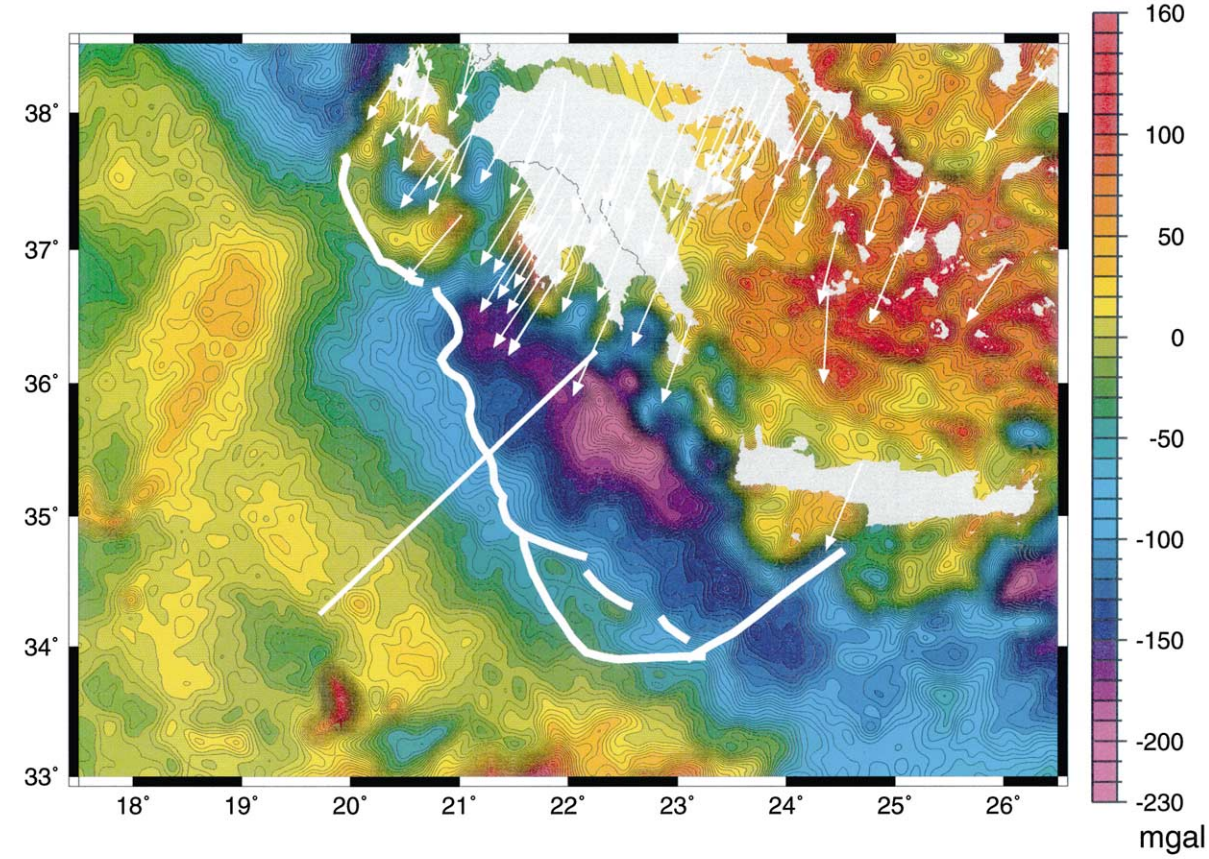

Fig. 4. Gravimetric map of the Ionian sea area deduced from satellite altimetry (Smith and Sandwell, 1997). The prism–backstop boundary is also shown. The arrows show the motion of the Aegean block with respect to the African plate as deduced from space geodesy (modified from Le Pichon et al., 1995 and Kahle et al., 1995).

form fault, the convergence velocity with respect to Eurasia becomes small and consequently the velocity with respect to Africa becomes close to that of the Eurasia/Africa convergence velocity (slightly less than 10 mm/yr).

The large gravity and topographic troughs to the south of 37°N coincide with the zone of fast Aegea/Africa convergence velocities whereas the gravity high and the more complex topography to the north occur opposite slower convergence velocities. We propose that this difference reflects a change from shortening by subduction of the oceanic lithosphere to the south, to mostly distributed shortening within the lower continental margin to the north, explaining the absence of intermediate type earthquakes to the north (see for example Hatzfeld, 1994 and Hatzfeld et al., 1995).

3. The migration of the Hellenic nappes from the Aegean continent to the adjacent Mediterranean basin

In a synthesis paper, Aubouin et al. (1976) noted that the geometry of the Hellenic nappes stack on land implied that their last phase of seaward migration extended at sea and that the front of the nappes was situated at least 100–150 km off the southern coast of Crete to the southwest, but much closer to the coast off Central Peloponne-

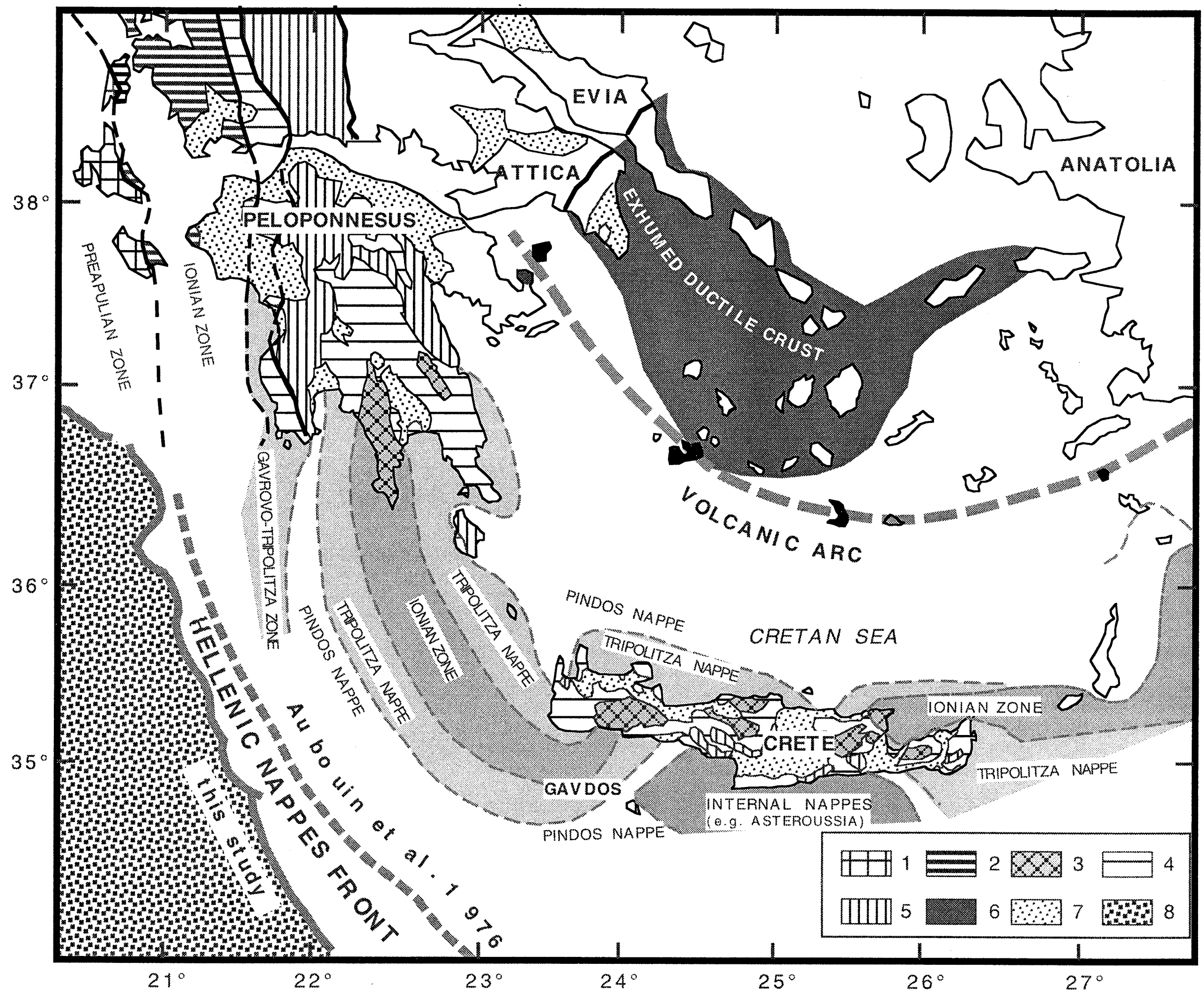

Fig. 5. Simplified structural map of the Aegean arc showing the distribution of the structural units (Hellenic thrust sheets from a synthesis by Aubouin et al., 1976), the exhumed metamorphic core complexes (after Jolivet et al., 1996) and the nappes leading edge according to Aubouin et al. (1976) compared to the backstop leading edge. 1: Preapulian Zone (mostly platform limestones); 2: Ionian Zone (basinal sequences); 3: 'Plattenkalk' series interpreted as metamorphosed Ionian series; 4: Gavrovo–Tripolitsa Zone (mostly platform limestones); 5: Pindos Zone (radiolarian bearing basinal sequences); 6: metamorphic core complexes; 7: Neogene sedimentary basins; 8: Mediterranean Ridge accretionary complex.

sus, to the northeast; thus, in their interpretation, the maximum displacement of the front of the nappes would increase from north to south (see Fig. 5). Bonneau (1973) later showed that the nappes he studied in Crete had essentially reached their present situation with respect to the underlying basement some time between the lower Oligocene and the middle Miocene and Thiébault (1982) further showed that the Gavrovo-Tripolitza nappe, that is the outermost nappe outcropping within Peloponnesus (see Fig. 5), reached its present location with respect to the underlying basement between the end of Oligocene (25 Ma) and the beginning of the Burdigalian (20 Ma). The outward limit Aubouin et al. (1976) proposed at the time for the front of the Hellenic nappes approximately coincides with the outward limit of the backstop. In Fig. 5, we have extended the front of the nappes northward toward the Ionian islands where the Ionian thrust is now situated. It has been shown since that the displacement of the nappes at the latitude of northern Peloponnesus is indeed much less than at the level of southern Peloponnesus and Crete (Thiébault, 1982).

Although the arguments of Aubouin et al. (1976) assumed cylindrical continuity and ignored the subsequent Miocene phases of gravitational extension and collapse (e.g. Lister et al., 1984; Buick, 1991; Gautier et al., 1990, Lee and Lister, 1992; Dinter and Royden, 1993; Jolivet et al., 1994; Jolivet et al., 1996; Avigad et al., 1997), the similarity in geometry between the outward limit of the nappes proposed by Aubouin et al. (1976) and the outward limit of the backstop as obtained from the new bathymetric data leads us to assume that they are identical.

We stated earlier that the backstop has a thickness of 6–8 km and a rather high seismic velocity of 4.8–6.5 km/s. This thickness is compatible with the combined thickness given by Thiébault (1982) for the Pindus and Gavrovo-Tripolitza nappes in Peloponnesus and the seismic velocities are also compatible with the lithologies of these nappes (indurated limestones to flysch series).

It is thus tempting to consider that the core of the Mediterranean Ridge backstop actually consists of the Hellenic nappes (probably mostly the Gavrovo-Tripolitza and Pindus nappes). Fig. 6 is a schematic diagram of the kinematics of emplacement of the nappes. We use the simplest pos-

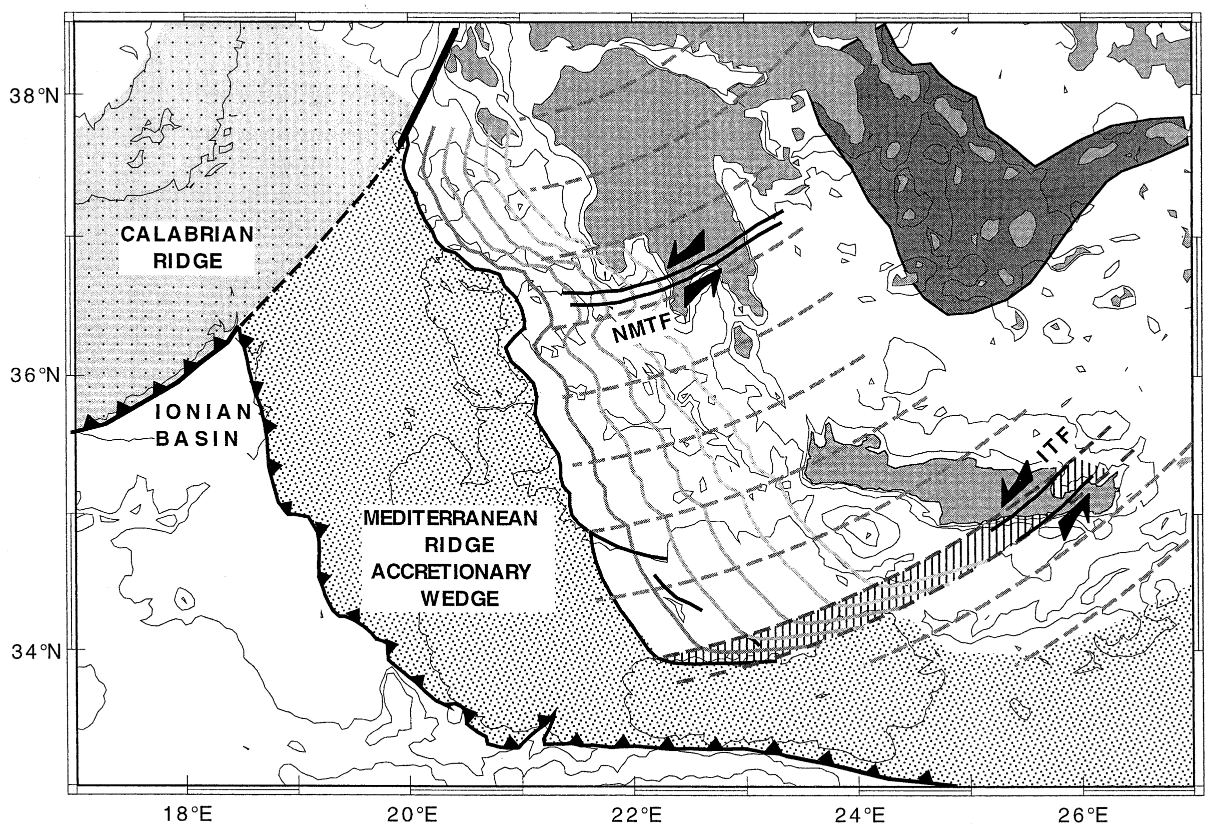

Fig. 6. Possible kinematics of the backstop emplacement in late middle Miocene. Limits of the core complexes are as in Fig. 5. ITF stands for Ierapetra Transverse Fault and NMTF for the North Mani Transverse Fault. The finite rotation angle is 15° and intermediate positions of the leading edge are shown every 3°.

sible description of this motion, that is a 15° rigid rotation with respect to the present position of the Aegean continent. The eulerian pole is located near 40.3°N, 20.5°E. Although it may seem unlikely that the motion of the nappes is approximately described by a rigid rotation, some additional support for the validity of this kinematic description may be obtained from the remarkably coherent paleomagnetic rotations during the last 5 Myr measured by Kissel and Laj (1988), or Kissel and Speranza (1995) on top of the nappes within the Ionian islands.

4. The Ierapetra and North Mani faults and the kinematics of emplacement of the nappes

We now discuss the existence of two major faults extending from the continent to the backstop that were active during the late middle Miocene, the Ierapetra fault zone in eastern Crete (Fig. 7) and the North Mani fault zone in southern Peloponnesus (Fig. 8). Fortuin and Peters (1984) proposed that the Ierapetra basin in eastern Crete is part of a probably left-lateral strike-slip zone, limited by the Ierapetra fault to the

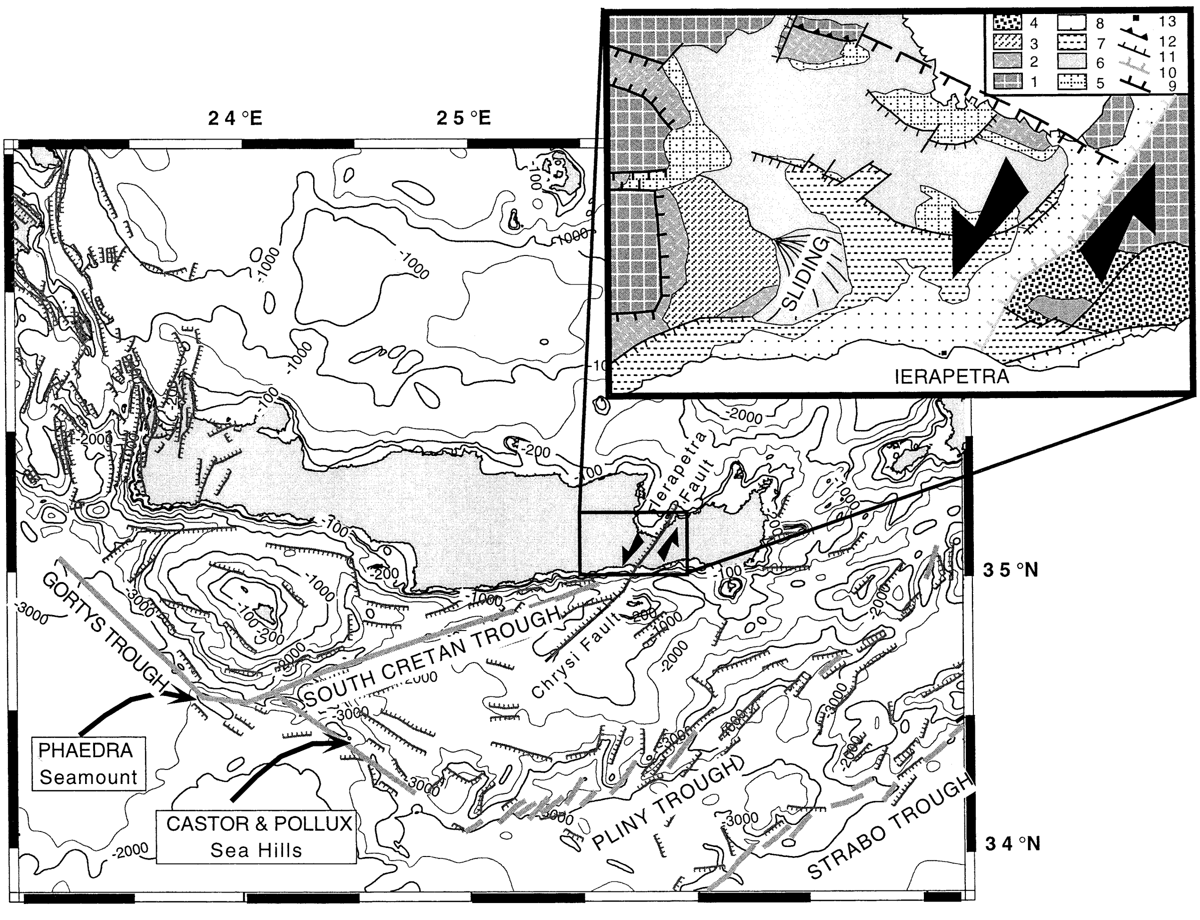

Fig. 7. Structural elements concerning the Ierapetra Transverse Fault system. Faults at sea are taken from Leite (1979), Angelier et al. (1982) and Nesteroff et al. (1977). Bold gray lines correspond to the deepest basins known as the 'Hellenic Trenches'. The inset shows the geological map of the Ierapetra area (from Fortuin and Peters, 1984). 1: Gavrovo–Tripolitsa nappe (mostly Mesozoic limestones); 2: internal nappes (e.g. Asteroussia); 3: Serravalian Mithi and Males formations; 4: Late Serravalian (east of the Ierapetra fault); 5 and 6: Late Serravalian Prina complex (west of the Ierapetra fault, contemporaneous of the strike-slip event); 7: Latest Serravalian and Tortonian; 8: post-Tortonian series; 9: late Miocene major normal faults; 10: Pliocene to Quaternary major normal faults; 11: minor normal faults; 12: synsedimentary late Miocene thrust fault; 13: Town location.

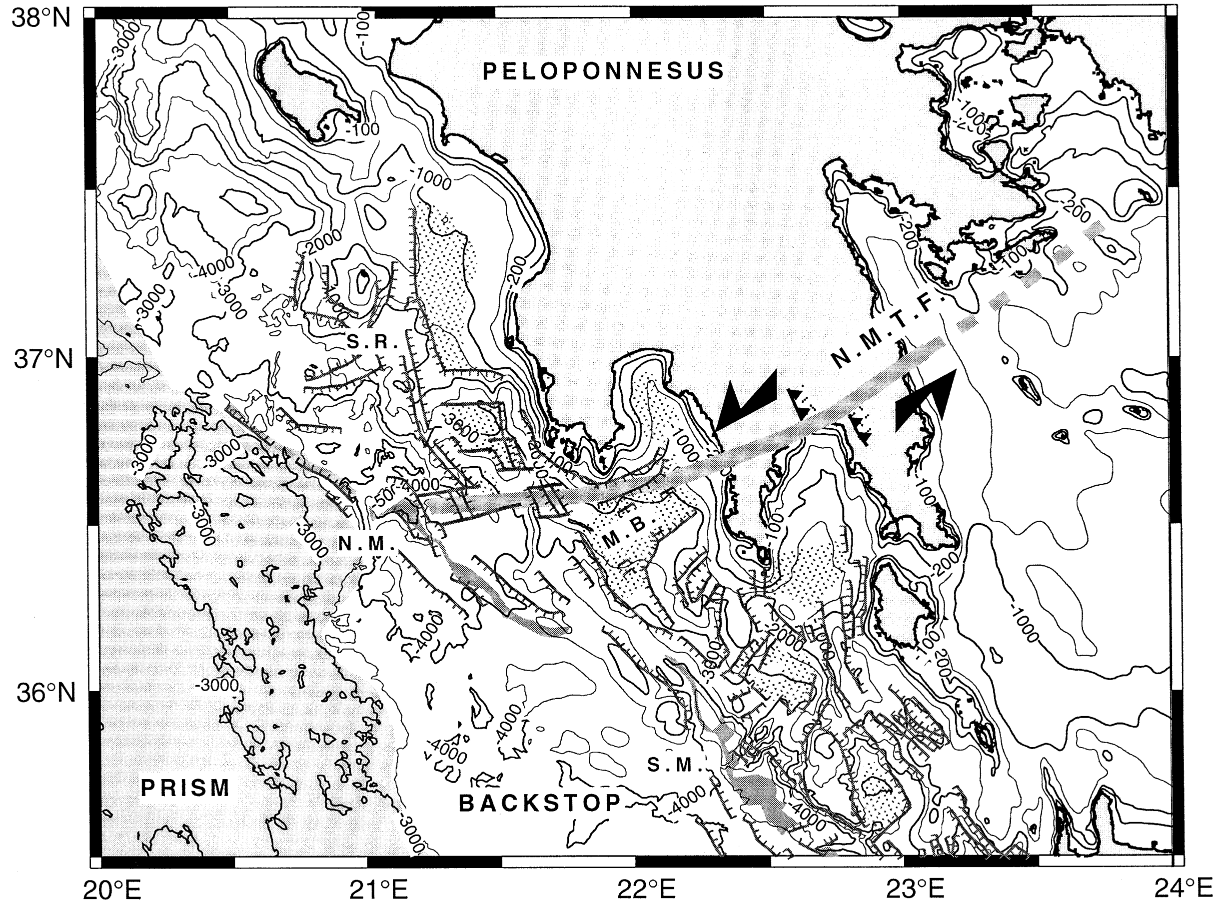

Fig. 8. Structural elements concerning the North Mani Transverse Fault (NMTF) modified from Lallemant (1984) and Lallemant et al. (1984). Faults at sea from Le Quellec et al. (1980). N.M.: North Matapan trough; S.M.: South Matapan trough; M.B.: Messenia Basin; dotted areas: Main Plio–Quaternary sedimentary basins.

east, which was initiated in Late Serravallian, about 13 Ma. They noted that there was no indication of post-Miocene lateral displacements but that this Ierapetra tectonic zone was reactivated as a prominent normal fault (see also Angelier, 1979; Angelier et al., 1982). They proposed further that the South Cretan Trough which ends near the Ierapetra basin is part of this NE–SW strike-slip zone (Fig. 7). However, Nesteroff et al. (1977) have proposed another offshore extension in the Chrysi fault which is in the exact prolongation of the onland Ierapetra fault. Considering that the South Cretan Trough is parallel to the Plio–Quaternary Pliny and Strabo troughs and significantly oblique to the Ierapetra fault (Fig. 7), we prefer this latter hypothesis. The recent left-lateral strike-slip deformation within the backstop would thus have a different direction from the late middle Miocene one, initiating fresh crustal faults within the backstop in Pliocene time, as shown since in Crete by Duermeijer et al. (1998).

Lallemant (1984) and Lallemant et al. (1984) described in southern Peloponnesus an E–W to ENE–WNW trending prominent left-lateral fault (the 'North Mani Transverse Fault' or NMTF) that cuts through the pile of nappes (Fig. 8). An offset of 10 km post-dates motion of the nappes and is older than upper Pliocene. They noted further that this fault zone extended offshore to the west where it marks the boundary between the North Matapan trough to the south and the Strophades Ridge to the north. The E–W trending North Matapan basin corresponds to an apparent left-lateral offset of the continental slope edge with a magnitude similar to the NMTF onland. One can also notice that this westward extension of the NMTF is actually made of several en échelon faults bounding NW–SE trending ridges and basins (Fig. 8). They might correspond to Riedel type left-lateral fractures related to the main strike-slip motion. There is no evidence for strike-slip motion along this fault since the late Pliocene. Instead, parts of this fault system acted as normal faults bounding prominent transverse basins as the offshore Messinia basin. Comparing this feature with the well dated Ierapetra fault in Crete and with other deformations in Peloponnesus, Lallemant et al. (1984) proposed a similar Late Serravalian age for both the Ierapetra fault and the NMTF. Lybéris and Lallemant (1985) proposed that the transition from subduction to the south to collision to the north occurred along the western extension of the North Mani fault zone.

Fig. 6 shows that the seaward limit of the backstop is also left-laterally offset along the western prolongations of these two fault zones. Actually, southeast of the Ierapetra fault, the backstop extent appears to be much more limited although its exact configuration is not yet known. Fig. 5 shows that the assumption of cylindricity of the nappes structure made by Aubouin et al. (1976) becomes quite tenuous east of Crete, and is probably not valid there, as most of the supposed eastward extension of the isopic zones lies under water. Thus, the Ierapetra fault probably limits to the southeast the zone of wide backstop.

Fig. 6 also shows that the two faults approximately follow small circles of the rotation describing the kinematics of the motion from the Aegean continent seaward of the large crustal body that now forms the backstop. Finally the age at which these faults zones were last active is late middle Miocene, that is the age attributed to the initiation of the present accretionary wedge system. Thus, these faults can be considered to have acted as the approximate equivalents of transform faults during the motion of the backstop crustal body that consequently must indeed have moved coherently.

5. Relationship between the outward motion of the Hellenic nappes and the contemporaneous Aegean extension

Since the work of Aubouin et al. (1976), it has been demonstrated that after the Paleogene phase of nappe stacking, the central Aegean has been the site of intense crustal extension since the beginning of Miocene and this extension has formed core complexes through uplift of the lower middle continental crust to the surface during the middle upper Miocene (Lister et al., 1984). Although the extension had clearly started by 20 Ma (Buick, 1991) and perhaps even late Oligocene (Gautier et al., 1990; Gautier et al., 1993), the peak of the uplift and presumably of the extension occurred between 12 and 10 Ma (Lee and Lister, 1992). Similar extension occurred during the same period within the northern Aegean (Dinter and Royden, 1993) and in southwestern Turkey (Hetzel et al., 1995a; Hetzel et al., 1995b). Jolivet et al. (1994) have proposed that the brittle extension that affects the Aegean since middle Miocene (Angelier et al., 1982; Mercier et al., 1979; Meulenkamp et al., 1988) is the direct continuation of this earlier ductile deformation. In Crete, the high pressure maximum, that presumably coincides with the maximum thickness of the nappes, before the extension started, occurred between 25 and 20 Ma (Jolivet et al., 1996). Jolivet et al. further showed that the ductile extensional decollement that led to the formation of the sea of Crete started to be active toward the north by 15–14 Ma. We have seen earlier that the nappes originating from the Aegean continent had probably reached their present location near the top of the Mediterranean Ridge by this time. This 20–15-Ma time span suggests a seaward migration velocity of the nappes of 20–30 mm/yr southwest of Crete.

We thus conclude that the upper middle Miocene migration of the nappes that formed the present Mediterranean backstop appears to be related to the end of this phase of continental collapse. It is remarkable that the central Aegean core complex is best developed opposite the zone of maximum seaward displacement of the Hellenic nappes with respect to the present Aegean continent(see Fig. 6) whereas this seaward displacement is minimal west of northern Greece and probably south of eastern Aegean where the core complex is least developed. It seems reasonable to assume that the present location of the core complexes were in late Oligocene to early Miocene the sites of maximum crustal thickness and highest topography, and that they became the sites of maximum crustal extension as reflected in the amount of seaward spreading of the Hellenic nappes.

6. The geological significance of the middle Miocene formation of a new Mediterranean accretionary complex

The present rate of subduction below Aegea is about four times larger than the rate of motion of Africa with respect to Eurasia. The larger rate is due for about one half to the Anatolia extrusion and for another half to the Aegean extension, taking the estimate of average rate of Aegean extension from the reconstruction of Le Pichon and Angelier (1981) and the rate of subduction from Le Pichon et al. (1995). As the westward extrusion of Anatolia did not start before 13 Ma at the earliest to 5 Ma at the latest (Sengör et al., 1985; Sengör, 1979, Le Pichon et al., 1995), the corresponding shortening rate between Aegea and Africa during lower to middle Miocene was about half the present one. And it was only one fourth of the present one (that is the Africa/Eurasia rate) prior to the gravity collapse of the Aegean continent, during Oligocene. Yet, at this time, crustal shortening occurred within Aegea which appeared to have been a large active crustal wedge (Avigad et al., 1997) and must have taken at least part of this shortening rate. Thus very little shortening was available for subduction during Oligocene. Rough estimates show that the present length of the seismic slab accounts for about 20 to at most 25 Myr of subduction.

The simplest interpretation then is that there was no subduction prior to early Miocene. The relatively slow shortening before that time was taken by internal deformation of Aegea. The initiation of the Aegean continent extension started probably during Aquitanian in central Aegea but only during Burdigalian in Crete (Jolivet et al., 1996). The collapse of the continent led part of the stack of nappes to move from the continent to the adjacent oceanic basin sometime during Burdigalian. A trench finally formed at the front of the advancing stack of nappes during upper middle Miocene. Once the subduction zone was fully formed, the roll-back of the trench accelerated the collapse of the continent that appeared to have been maximum near 12–10 Ma (Lee and Lister, 1992).

However seismic tomography has demonstrated the presence of a very long high velocity slab in the prolongation of the seismic slab (Spakman, 1990; Spakman et al., 1988, 1993) and has been interpreted as indicating a continuous subduction since at least Eocene time or even earlier. This very long length of the slab poses difficult problems that we cannot answer simply. The first difficulty concerns the question of the slow rate of shortening between Aegea and Africa during Oligocene and Eocene, as discussed above. What is the significance of a subduction zone in which the subduction rate is only a few mm/yr during at least 20 Myr? Should not the slab be verticalized and detached? The second difficulty concerns the location of the pre-middle Miocene subduction zone. The present accretionary complex was initiated in upper middle Miocene time and the crustal body that forms the backstop appears to have reached its present position during the same period. If a subduction zone already existed in Oligocene and early Miocene, where are the remains of the corresponding accretionary complex? Has it been destroyed by subcrustal erosion or overridden by the collapse of the Aegean continents? Is it then possible that the present continental margin is a recent one, possibly due to underplating and uplift of the arc (Le Pichon and Angelier, 1981) and (or) to subcrustal erosion of the present backstop?

Whether a subduction zone existed prior to

Miocene or not, it is difficult to understand why the present western continental margin is so steep and rectilinear. The rectilinearity suggests a recent control by a strike-slip fault. An observation that may favor this interpretation is the following one. The present Pliny and Strabo trenches to the south of Crete appear to be post-Messinian and perhaps post-3 Ma as massive gravity deposits of this age have been dredged on the summits of mounts Phaedra, Castor and Pollux which are now isolated within these troughs (Peters and Troelstra, 1984) (see Fig. 7). If these left-lateral Pliny and Strabo trenches are related to the extrusion of the Mediterranean Ridge due to its collision with the African continental margin since 3–5 Ma as proposed by Le Pichon et al. (1995), their Pliocene age can be understood. It is then possible that the steep rectilinear western Aegean continental margin also originated as a strike-slip fault (right lateral) facilitating the northward extrusion of the western Mediterranean Ridge and that it later localized normal motion, shaping the morphology of the present continental margin.

7. Conclusion

We have proposed that the present backstop had been formed in middle Miocene time by the outward migration of the Hellenic nappes between 20 and 15 Ma. The present accretionary wedge then would be post-middle Miocene. This interpretation is supported by the similarity in thickness and seismic velocity between the backstop and the Hellenic nappes in southern Peloponnesus. It is also supported by the fact the extension of the backstop away from the Aegean continental margin is maximal to the southwest where the displacement of the nappes with respect to their substratum is also maximal and becomes quite small near the Ionian islands where this displacement is small. We have not found a simple way to reconcile the present data with the existence of significant subduction below the Aegean continent during Eocene to lower Miocene time.

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