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Geodynamic connection between the indentation of Arabia and the Neogene tectonics of the central–eastern Mediterranean region

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ABSTRACT

The indentation of Arabia has influenced the post-Oligocene evolution of the eastern Mediterranean–European region, from Anatolia to the Carpathians and even the central Mediterranean area, since the late Miocene. The efficient transmission of compressional stress from the Arabia-Eurasia collision zone into such a broad area was facilitated by the Tethyan orogenic belt, developed during the Late Cretaceous–Paleogene closure of the northern Neotethyan Ocean, and in particular by its strong inner metamorphic core. The lateral displacement of this orogenic belt, due to the push of Arabia, was progressively accommodated by bowing and extrusion of those sectors that faced low-buoyancy lithospheric domains. This extrusion process first involved the outward expulsion of the Carpathian arc at the expense of the low-buoyancy Magura oceanic domain, and then the lateral escape of the Cycladic arc at the expense of the low-buoyancy Ionian-Levantine oceanic domain. The lateral migration of the Tethyan orogenic belt began to influence the deformation pattern of the central Mediterranean region around the late Miocene, when its Cycladic-Pelagonian sector welded into the southern Adriatic continental domain. Under the westward push of the Tethyan belt, the Adria plate decoupled from Africa and from its Padanian protuberance and rotated clockwise at the expense of the western Apulian low-buoyancy zone. The available geological evidence in the study area suggests that volcanic activity was spatially associated with extension in the upper crust. We discuss the possible implications of this inferred geodynamic connection between the Arabian indentation and regional deformation patterns for the temporal evolution of deep tectonic processes in the central–eastern Mediterranean.

Keywords: Mediterranean geodynamics, plate tectonics, continental deformation, back-arc basins

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INTRODUCTION

It is widely recognized that Anatolia (Fig. 1) has extruded westward in response to the indentation of the Arabian promontory of Africa (e.g., McKenzie, 1978; Dewey and Şengör, 1979; Hempton, 1987). However, it is a topic of debate as to whether and how this extrusion influenced the Neogene deformation in the western Anatolian, Aegean, and peri-Adriatic regions. While this influence is recognized by some authors (Tapponnier, 1977; Şengör et al., 1985; Mantovani et al., 1997a, 2002; Yılmaz et al., 2000; Mantovani, 2005), some researchers (e.g., Le Pichon, 1982; Meulenkamp et al., 1988; Seyitoğlu and Scott, 1996; Hatzfeld et al., 1997; Gautier et al., 1999; Jolivet, 2001) invoke alternative driving mechanisms, such as gravitational spreading and slab-pull forces.

In this article we present extensive evidence and arguments in support of the hypothesis that the indentation of Arabia has been a primary driving force for the observed deformation pattern in the eastern Mediterranean region since the Oligocene and even in the central Mediterranean since the late Miocene. The proposed evolutionary history is based on the hypothesis that the convergence of the confining plates, Africa, Arabia, and Eurasia, has mainly been accommodated by lateral escape of buoyant orogenic bodies at the expense of the low-buoyancy oceanic zones (Ionian, Levantine, western Apulian, and Magura; see Fig. 2). The importance of extrusion tectonics in the Mediterranean area has already been suggested by a number of authors (e.g., Tapponnier, 1977; Boccaletti et al., 1982; Mantovani et al., 1997a, 2002, 2006; Mantovani, 2005). In this study, we point out that to fully understand the observed deformation pattern in the study area it is necessary to consider the crucial role played by the long orogenic belt created by the closure of the northern Neotethyan Ocean, hereafter cited as the Tethyan belt. The high strength and cohesion of this body, in particular of its metamorphic core, allowed it to efficiently transmit stresses from the Arabia-Eurasia collision zone to the surrounding regions and to maintain its continuity, notwithstanding the strong deformation it has undergone during its evolution.

A key tectonic process widely recognized in the Mediterranean area is the development of trench arc–back-arc**[AQ1]** systems that led to the opening of major Neogene basins, such as the Balearic, Tyrrhenian, Aegean, and Pannonian basins, and to the strong deformation of the pre-Neogene orogenic belts. We argue that such a process develops where a buoyant orogenic belt, facing a low-buoyancy domain, is stressed by belt-parallel compression. The implications of this hypothesis for the opening of the Tyrrhenian, Aegean, and Pannonian back-arc basins are discussed later, whereas the implications for the opening of the Balearic basin have been discussed in Mantovani (2005). In this article we also discuss the possible connection between the extensional deformation predicted by the proposed geodynamic evolution and the time-space distribution of volcanic activity, in particular related to subduction-related magmatism. Finally, we evaluate the arguments and evidence for various geodynamic models so far proposed in the literature.

EVOLUTIONARY PATTERN

Oligocene Paleogeographic Setting

Figure 2 shows the tectonic setting of the central–eastern Mediterranean area we adopt as the starting point of our evolutionary reconstruction. This paleogeographic configuration is compatible with those proposed by most previous reconstructions (e.g., Boccaletti and Guazzone, 1974; Burchfiel, 1980; Rehault et al., 1984; Burtman, 1986; Dercourt et al., 1986; Royden and Baldi, 1988; Finetti et al., 1988, 2001; Finetti, 2004).

The African and Eurasian domains were separated by a rather heterogeneous structure, including a continental domain (the Adriatic); relatively large oceanic zones with low-buoyancy lithosphere, such as the Ionian, Levantine, western Apulian, and Magura (e.g., Royden and Baldi, 1988; Finetti et al., 1996, 2001; Chalot-Prat and Girbacea, 2000; Csontos and Vörös, 2004); and a long orogenic system generated by the consumption of the northern Neotethys domain since the Cretaceous and accretion of continental fragments situated between small oceanic basins (e.g., Şengör and Yılmaz, 1981; Dercourt et al., 1986; Robertson et al., 1996; Gessner et al., 2001; Ring and Layer, 2003; Golonka, 2004).

The Tethyan domain consisted of a series of oceanic basins and continental fragments derived by the early Mesozoic rifting of the northern margin of Gondwanaland. The consumption of oceanic domains generated the main ophiolitic belts (e.g., Dilek and Moores, 1990; Dilek et al., 1999; Robertson, 2002), whereas underthrusting processes at continental collision boundaries generated the main metamorphic belts (Ricou et al., 1998; Okay and Tüysüz, 1999; Gessner et al., 2001; Whitney et al., 2001). As pointed out by a number of researchers (e.g., Brunn, 1976; Boccaletti and Dainelli, 1982; Burtman, 1986; Royden and Burchfiel, 1989; Okay and Tüysüz, 1999; Robertson, 2002; Ring and Layer, 2003), this orogenic system included (Fig. 2) an inner core (light and dark violet belts in Fig. 2) consisting of oceanic remnants and crystalline massifs (Pelagonian-Cyclades-Anatolian) and two external accretionary chains, one with European affinity (Carpathians, Balkanides, and Pontides) and one with African affinity (Dinarides, Hellenides, and Taurides).

The geodynamic processes that turned the initial structural setting (Fig. 2) into the present one (see Fig. 1 and Fig. 6B later in this chapter) have involved the tectonic transport and strong distortion of orogenic belts and the formation of basins in the wake of the migrating or bending arcs. This deformation pattern is generally known as the generation of a trench arc–back-arc system, and the major effects of its development are clearly recognized in the Carpatho-Pannonian (e.g., Royden, 1993a,b; Csontos and Vörös, 2004), Tyrrhenian-Apennines (e.g., Sartori, 1990; Sartori and Capozzi, 1998), and Aegean (e.g., Le Pichon

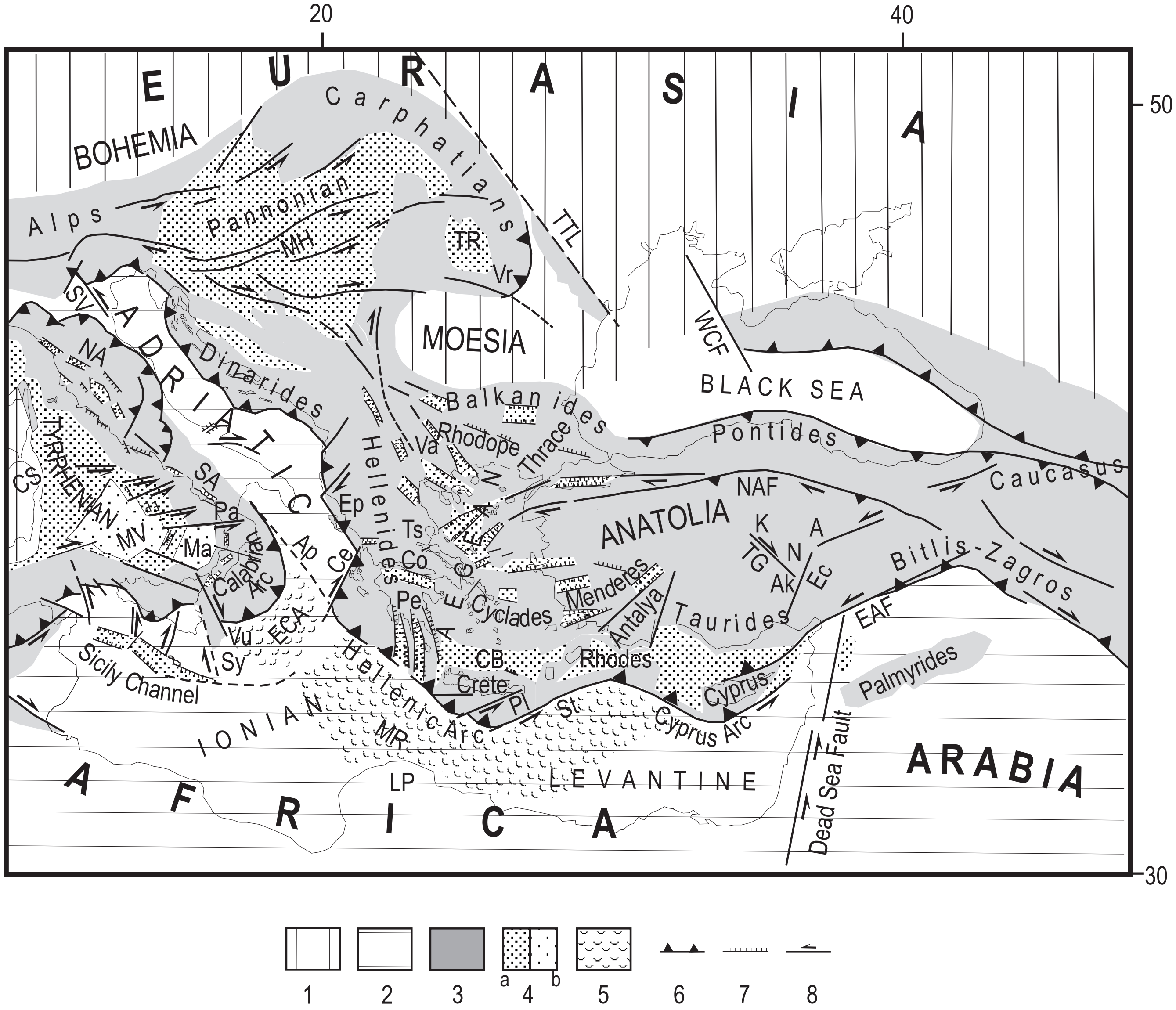


Figure 1. The present tectonic setting in the central–eastern Mediterranean region. Key to numbering of symbols and shading: (1) Eurasian domain; (2) African domain; (3) orogenic belts; (4) zones affected by moderate (a) or intense (b) crustal thinning; (5) Mediterranean Ridge (MR) and external Calabrian arc (ECA) accretionary complexes; (6, 7, 8) compressional, tensional, and strike-slip features, respectively. A—Akdag massif; Ak—Aksaray massif; Ap—Apulian escarpment; CB—Cretan basin; Ce—Cephalonia fault system; Co—Corinth trough; CS—Corsica-Sardinia microplate; EAF—east Anatolian fault; ECA—external Calabrian arc; Ec—Ecemis fault; Ep—Epirus; K—Kirsehir massif; LP—Libyan promontory; Ma—Marsili basin; MH—mid-Hungarian fault system; MR—Mediterranean Ridge; MV—Magnaghi-Vavilov basin; N—Nigde massif; NA—northern Apennines; NAF—north Anatolian fault system; Pa—Palinuro fault; Pe—Peloponnesus; Pl—Pliny fault; SA—southern Apennines; St—Strabo fault; SV—Schio-Vicenza fault; Sy—Syracuse escarpment; TG—Tuz-Golu fault; TR—Transylvanian basin; Ts—Thessaly; TTL—Tornquist-Teisseyre Line; Va—Vardar zone; Vr—Vrancea zone; Vu—Vulcano fault; WCF—west Crimean fault.

and Angelier, 1979; Mercier et al., 1987; Armijo et al., 1996) regions, as well as in the western Mediterranean area (Rehault et al., 1984; Dercourt et al., 1986). Later we discuss the mechanisms for such tectonic process, as well as the proposed dynamics and the general conceptual framework on which our reconstruction is based.

Driving Forces and Tectonophysical Concepts

In our geodynamic interpretation, we assume that the observed deformation pattern in the Mediterranean area has been driven mainly by the convergence of the confining plates, Africa, Arabia, and Eurasia. Obviously, we do not neglect the

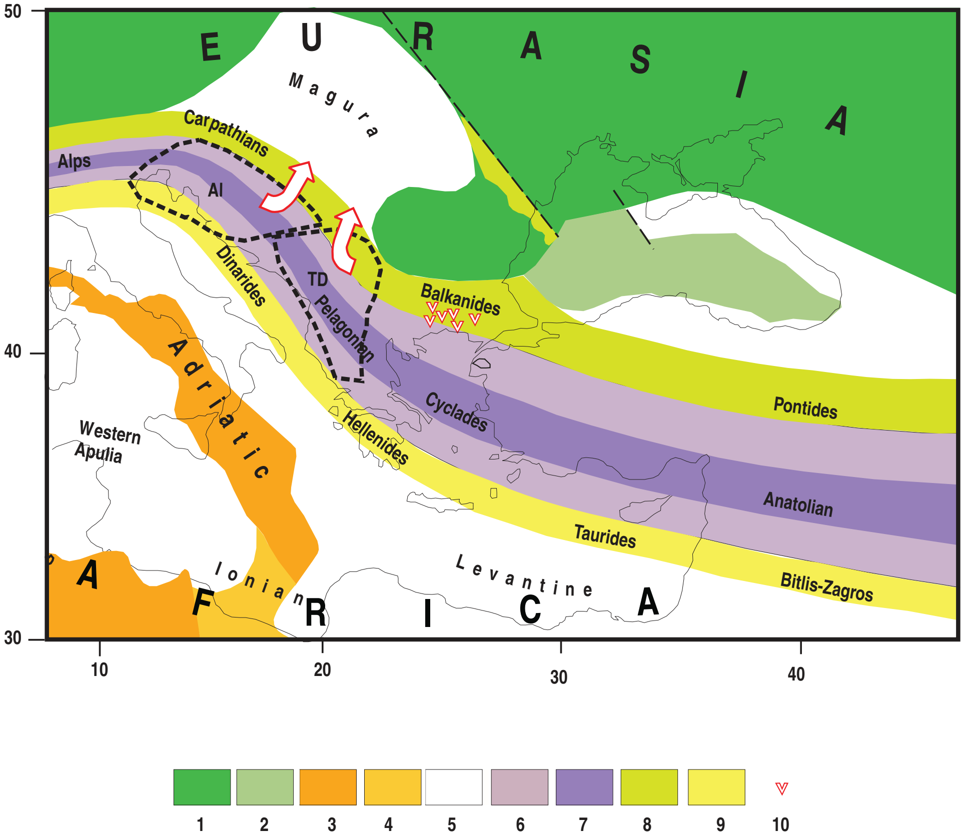


Figure 2. Tentative reconstruction of the Oligocene paleogeography of the eastern Mediterranean region. Key to numbering of shading: (1, 2) Continental and thinned Eurasian domains, respectively, shown with the present extension; (3, 4) continental and thinned African and Apulian domains, respectively, shown with the present extension; (5) lithospheric domains that have been consumed during evolution; (6, 7, 8, 9) Tethyan belt, constituted by an inner core, with ophiolitic and metamorphic zones (6), crystalline massifs (7), and external accretionary belts, with European (8) and African (9) affinity; (10) volcanic zones. Al—Alcapa block; TD—Tisza-Dacia block. The white arrows indicate the opposite rotations of the Alcapa and Tisza-Dacia blocks during their subsequent evolution. The present geographical contours (thin black lines) and paleoposition of the African coastline (thick black lines) are shown for reference. Other symbols are as in Figure 1. Tectonic features are not displayed in this figure; only the initial configuration of main structural elements is depicted.

role of gravitational forces in structural settings where density contrasts are involved, but we do not share the opinion of other authors who recognize a primary role of such forces in crustal tectonic processes, as implied, for instance, by the slab-pull and gravitational spreading models. Remarks about this problem in the study area are given in the section headed “Discussion.”

It is widely recognized that plate convergence can be accommodated by various kinds of shortening processes, such as subduction of lithosphere, crustal thickening, and lateral escape of buoyant crustal wedges at the expense of low-buoyancy lithosphere. Physics predicts that the time-space distribution of shortening is controlled by the minimum-work principle, i.e., the need to minimize the total work of horizontal, kinematically induced forces against any kind of resistance, such as buoyancy, friction, and viscous resistance of the mantle (e.g., Sleep et al., 1979; Molnar and Lyon-Caen, 1988; Masek and Duncan, 1998).

In subduction processes the amount of work done by driving forces mostly depends on the structural and rheological properties of the subducting lithosphere, for these features control the volume of buoyant crustal material that must be scraped off the descending lithosphere and accumulated in the accretionary prism. The sinking of the nonbuoyant part of the lithosphere occurs at the expense of the potential gravitational energy previously stored in the system, and thus it requires much less work from horizontal driving forces.

The considerations previously stated imply that the consumption of cold, dense oceanic lithosphere, not including intracrustal asthenospheric layers (Ranalli and Murphy, 1987; Lobkovsky and Kerchman, 1991), is the most efficient subduction process. In fact, such lithosphere is denser than the surrounding mantle (Cloos, 1993) and thus subducts as a whole, leaving only very limited amounts of accreted material (mostly sedimentary rocks) in the trench zone, as occurred, for instance, in the Calabrian and Hellenic (Mediterranean Ridge) arcs.

When the subducting lithosphere has a continental character instead, decoupling may occur between the nonbuoyant mantle part of the lithosphere and the buoyant crust. In this case, the amount of material accumulated in the trench zone increases considerably, as well as the work needed for fracturing and imbricating brittle upper crustal slivers and for emplacing ductile lower crustal material into or under the crust of the overriding plate (e.g., Van den Beukel, 1992; Meissner and Mooney, 1998). Thus, underthrusting of thick continental crust (>30 km) may lead to the termination of the subduction process.

Rheological profiles reconstructed for the various structural provinces of the Mediterranean area (Viti et al., 1997) indicate that lithosphere with no ductile decoupling layers exists only in the Ionian and Levantine old oceanic domains, whereas a ductile lower crust is present in most of the Mediterranean zones. Therefore, one can reasonably expect that the consumption of the low-buoyancy oceanic domains, which were left in the Mediterranean region after the Eocene (western Apulian, Ionian, Levantine and Magura; see Fig. 2), was the preferred shortening process for accommodating the convergence of the confining plates.

Once a stationary configuration of shortening processes is reached in the system, controlled by the boundary conditions and by the properties of the structures involved, it lasts until a significant change of dynamic conditions occurs at one or more of the ongoing consuming boundaries. This change may occur, for instance, when continental crust arrives at a consuming boundary, causing a considerable increase of buoyancy forces (e.g., Shemenda, 1993; Boutelier et al., 2004, and references therein).

The tectonic reorganization that follows this event, aimed at finding a new minimum work configuration, depends on the nature of the lithosphere in the collision zone and surroundings. If low-buoyancy oceanic lithosphere is not present in the near surroundings, the subduction of thicker lithosphere persists, in spite of the increased resistance. Of course, in this context the amount of buoyant crustal material scraped off the descending lithosphere and accumulated in the trench zone and below the upper crust of the overriding plate increases significantly, causing crustal thickening and uplift in the collisional boundary (e.g., Zhao and Morgan, 1985; Westaway, 1995; Meissner and Mooney, 1998). This process can continue for a long time, as occurred in the India-Eurasia boundary, with the formation of large and elevated accretionary belts. If, however, low-buoyancy lithosphere is present somewhere in the surrounding regions, the previously described process may be interrupted by the formation of major fault systems, which behave as lateral guides for the extrusion of buoyant orogenic crustal wedges from the collision zone toward the weak lateral boundary (e.g., Tapponnier and Molnar, 1976; Mantovani et al., 2001a; Mantovani, 2005). This kind of solution can be recognized in the lateral extrusion of the Anatolian wedge during the early Pliocene, when the north and east Anatolian fault systems allowed the westward escape of Anatolia, after an uplift of ~2 km of eastern Anatolia (Şengör and Kidd, 1979; Dewey et al., 1986; Yılmaz et al., 1987; Barka, 1992). Escape of wedges allows stress release in front of the indenter (e.g., Hubert-Ferrari et al., 2003), and then the initial extrusion generally coincides with the end of accretionary activity (suture) at the old consuming boundary. Gravitational spreading of crustal material away from the most uplifted zones may also contribute to the lateral escape of crustal wedges toward thinner, possibly oceanic, adjacent structures (e.g., Dewey et al., 1986; Meissner and Mooney, 1998).

In line with the previously mentioned considerations, we use the term *collision* to identify the beginning of interaction between two converging continental domains and the term *suture* to identify the end of accretionary activity and welding of the colliding blocks.

A major problem in the Mediterranean region is finding an explanation for the generation of trench arc–back-arc systems, because many examples of this phenomenon are recognized throughout this region. Mantovani et al. (1997a, 2001a, 2002) suggested that this phenomenon may occur in places where a segment of an orogenic belt is stressed by belt-parallel compression, which causes the lateral extrusion and bowing of the arc and its partial separation from the overriding plate (Fig. 3). This separation is accommodated by crustal thinning in the back-arc zone. Simultaneously, the extruding buoyant wedges load the adjacent subducting lithosphere, causing its sinking and progressive retreating of the trench zone.

The geodynamic context that produces belt-parallel compression in the arc may be quite variable from case to case. Most commonly, this phenomenon occurs when a buoyant rigid structure enters a sector of the active margin with an oblique angle to the trench, as tentatively sketched in Figure 3. When this condition is not fulfilled—as occurred, for instance, in the India-Eurasia, Arabia-Eurasia, Adriatic-Eurasia, and South America–Nazca boundary zones, where the direction of plate convergence has been almost perpendicular to the orogenic belt and/or low-buoyancy lithosphere has been lacking in the surrounding region—neither the outward extrusion and bending of the arc nor the back-arc extension develops (Mantovani et al., 2001a). The proposed mechanism can develop only if the buoyancy of the stressed belt is significantly higher than that of the lithospheric domain lying in front of it. Hence, for instance, the extrusion of the arc is strongly favored when it faces old oceanic lithosphere, which is presumably characterized by negative buoyancy (Cloos, 1993).

The occurrence of the proposed mechanism is also conditioned by the activation of few major fault systems in the stressed belt, which allows the extruding body to fragment in only a few relatively large crustal wedges, sliding and rotating each other (Fig. 3). The extrusion of highly fragmented material, for instance, would tend to occupy the entire space available (e.g., Sornette et al., 1993), preventing any separation of the arc from the overriding plate and, consequently, the development of back-arc extension.

The physical plausibility of extrusion processes and their possible importance in the generation of back-arc basins have been quantitatively demonstrated by a number of authors (e.g., Tapponnier et al., 1982; Peltzer and Tapponnier, 1988; Ratschbacher et al., 1991; Faccenna et al., 1996; Lavé et al., 1996; Mantovani et al., 2000a,b; 2001b; Hubert-Ferrari et al., 2003; Seyferth and Henk, 2004; Viti et al., 2004). In what follows, we argue that the conditions required for the occurrence of the previously described mechanism can be recognized in the central–eastern Mediterranean zones where trench arc–back-arc systems have developed (Carpatho-Pannonian, Apennines-Tyrrhenian, and Hellenic-Aegean systems).

We wish to point out that the proposed paleogeographic maps do not aim at reconstructing local deformation patterns; they only provide tentative geometries and paleopositions of major plates, microplates, and orogenic wedges. The evidence and arguments supporting the kinematics of Africa and Arabia with respect to Eurasia that we have adopted in our evolution-

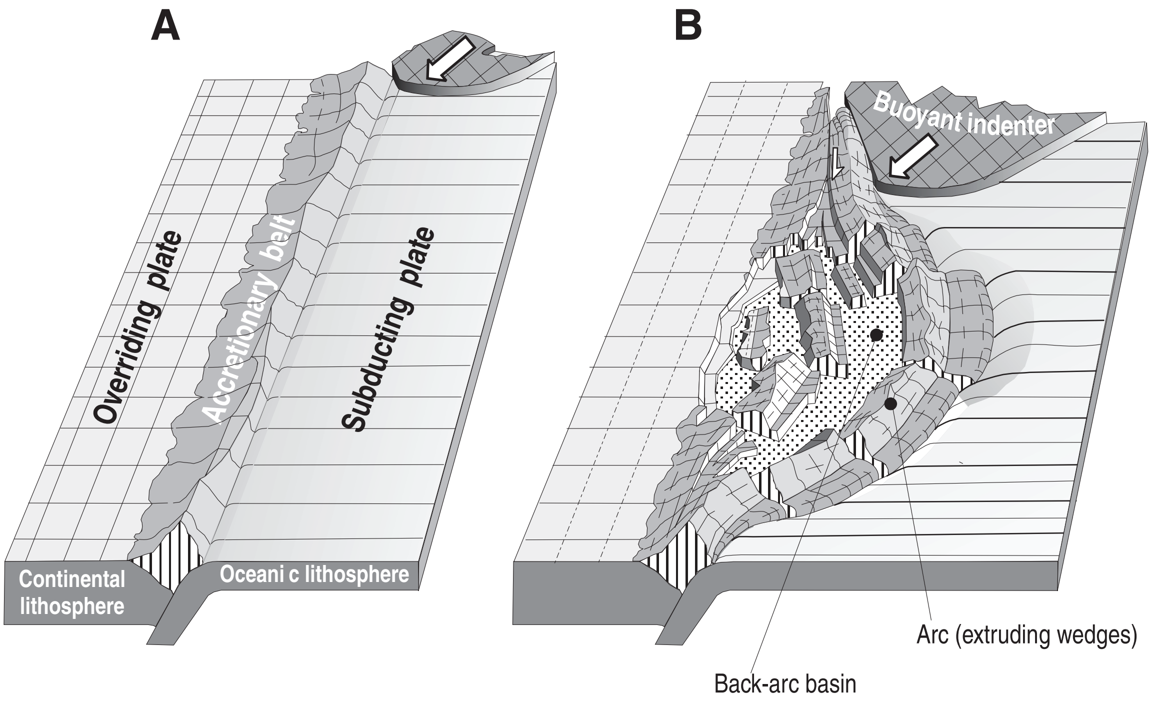


Figure 3. Proposed driving mechanism of a trench arc–back-arc system. (A) The structural or tectonic setting that may precede the proposed process. (B) A belt-parallel push, driven by the oblique indentation of a buoyant block, is accommodated by the lateral expulsion of crustal wedges, resulting in an outward bowing of the arc at the expense of the adjacent low-buoyancy oceanic lithosphere. The divergence between the bowing or extruding arc and the overriding plate is accommodated by crustal stretching in the back-arc zone (stippled area), within a limited sector of the boundary, while the outward migration of wedges causes subduction of the oceanic lithosphere and consequent accretionary activity along the arc front.

ary reconstruction are described by Albarello et al. (1995), Mantovani et al. (1997a, 2002), and Mantovani (2005).

Oligocene to Late Miocene

The observed deformation pattern in this evolutionary phase (Fig. 4A) involves a progressive outward migration and strong bowing of the Carpathian arc (confined by the rigid Bohemian and Moesian buttresses) at the expense of the low-buoyancy Magura oceanic domain (e.g., Burchfiel, 1980; Burtman, 1986; Royden and Burchfiel, 1989; Csontos and Nagymarosy, 1998; Fodor et al., 1998; Seghedi et al., 2004). In the internal part of this migrating arc, transtensional tectonics occurred along major fault systems (i.e., the mid-Hungarian fault; see Figure 4A and B). This kind of activity mainly developed in the northern part of the Pannonian area during the Oligocene–early Miocene and in the southern Pannonian region in the middle–late Miocene (e.g., Royden and Burchfiel, 1989; Royden, 1993a,b; Fodor et al., 1998; Hippolyte et al., 1999; Seghedi et al., 2004).

The deformation pattern typical of trench arc–back-arc systems, i.e., bowing of the arc with development of accretionary activity along its external front and extensional tectonics in its internal zone, is also recognized in the Aegean arc. Crustal extension affected the central–northern Aegean and northwestern Anatolian regions, as indicated by geological and volcanological evidence (e.g., Fytikas et al., 1984; Mercier et al., 1989; Papadopoulos, 1989; Seyitoğlu and Scott; 1996; Burchfiel et al., 2000; Okay and Satir, 2000). Thrusting is evidenced by seismic surveys along the outer front of the Aegean arc (e.g., Finetti, 1976; Mascle et al., 1999).

Later we discuss the idea that the previously mentioned deformation can be explained as a side effect of the push of the continental Arabian indenter after the complete consumption of oceanic lithosphere along its northern boundary (the Bitlis zone). This driving mechanism caused the decoupling between the Anatolian and the Iranian sectors of the Tethyan belt (Fig. 4). The inferred decoupling was facilitated by a system of northwest-oriented dextral transpressional faults. Evidence in support of this faulting is provided by a number of authors (Dercourt et al., 1986; Zonenshain et al., 1990; Andrieux et al., 1995; Golonka, 2004). Once decoupled, eastern Anatolia moved northwestward (Fig. 4), accommodated by shortening in the Caucasus

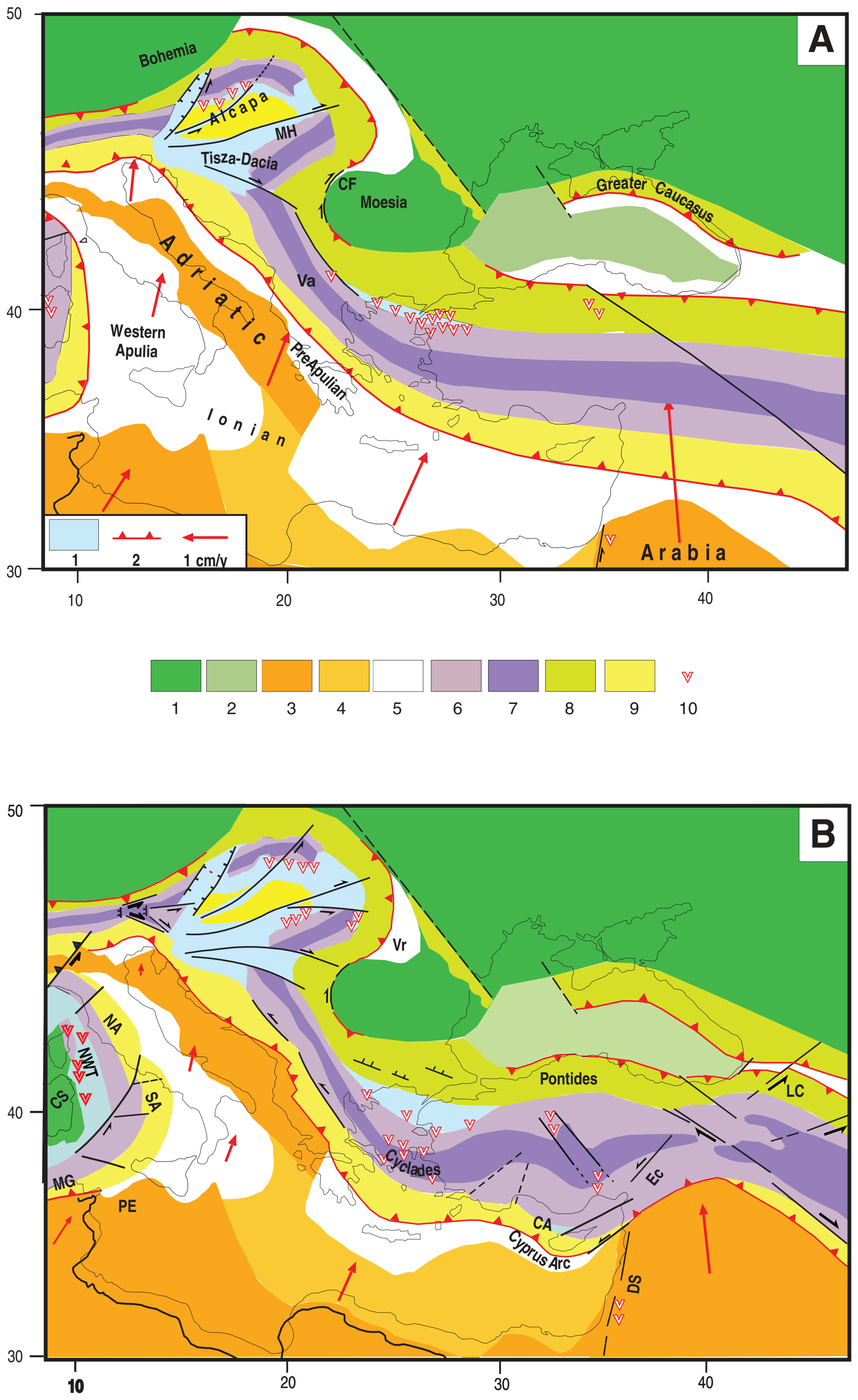


Figure 4. (A) The early Miocene paleogeographic setting. The blue areas depict the zones affected by moderate crustal thinning. CF—Cerna fault; MH—mid-Hungarian fault system; Va—Vardar zone. (B) The late Miocene paleogeographic setting. CA—Cilicia-Adana trough; CS—Corsica-Sardinia microplate; DS—Dead Sea fault system; Ec—Ecemis fault; LC—Lesser Caucasus; NA—northern Apennines; NWT—northwestern Tyrrhenian basin; PE—Pelagian zone; SA—southern Apennines; Vr—Vrancea zone. The red arrows indicate motion rates (scale in the inset) of major plates with respect to a Eurasian reference frame (Albarello et al., 1995; Mantovani et al., 1997a, 2002, 2004). Symbols and shading as in Figures 1, 2, and 4A. See text for explanations.

and the Pontides (Dercourt et al., 1986; Zonenshain and Le Pichon, 1986; Zonenshain et al., 1990). Contemporaneously, the Anatolian-Cycladic-Pelagonian sector of the Tethyan belt transmitted the push of eastern Anatolia to the Carpathian arc, which underwent outward extrusion at the expense of the Magura low-buoyancy domain. This last process caused the decoupling of the Tethyan belt from the Moesian block by a dextral shear zone, including the Cerna fault, recognized at the northwestern margin of Moesia (e.g., Schmid et al., 1998; Matenco and Schmid, 1999). While transmitting stress in the framework of the previously mentioned mechanism, the Cycladic arc underwent southward bowing at the expense of the adjacent Ionian-Levantine low-buoyancy domain (Fig. 4B). Due to this bowing, the arc partly diverged from the Rhodope-Balkan zone, causing extensional deformation and abundant volcanic activity in the northern Aegean and western Anatolia during the early–middle Miocene (Fytikas et al., 1984; Mercier et al., 1989; Papadopoulos, 1989; Wilson and Bianchini, 1999; Yılmaz et al., 2000).

A major role in the deformation of the Carpatho-Pannonian system was played by the relative rotation of crustal wedges, such as the Alcapa and Tisza-Dacia, as tentatively illustrated in Figures 2 and 4A and B. The mechanism we propose (Fig. 3) for the generation of the Carpathian and Cycladic arcs can explain the simultaneous occurrence of thrusting along their external fronts and transtensional tectonics in the internal zones, the Pannonian and northern Aegean regions, respectively. In the Pannonian area, crustal extension developed mainly along releasing bends of major shear zones (the mid-Hungarian fault) and in the zones where blocks underwent relative displacement and rotation (e.g., Csontos and Nagymarosy, 1998; Hyppolyte et al., 1999; Csontos and Vörös, 2004; Seghedi et al., 2004).

As the Magura low-buoyancy zone between the extruding Carpathian wedges and the continental Eurasian domain was almost completely consumed (Hyppolyte et al., 1999; Seghedi et al., 2004), all processes connected with this extrusion, such as accretionary activity on the external front and tensional or transtensional deformation in the internal Pannonian basin, underwent a considerable slowdown. Tectonic activity has continued only in the southeastern corner of this arc, the Vrancea zone (Hyppolyte, 2002; Seghedi et al., 2004).

The present morphology of the Cyprus arc developed around the late Miocene (e.g., Kempler and Ben Avraham, 1987; Kempler and Garfunkel, 1994; Robertson, 2000), as suggested by the timing of thrusting and extensional deformation, which occurred, respectively, along the outer and inner sides of this arc. This trench arc–back-arc system might have been developed by a mechanism similar to that sketched in Figure 3. The indentation of Arabia may have caused the westward escape of a narrow wedge of the Tauride belt, inducing the outward bowing of the Cyprus arc (Fig. 4B). The separation of this arc from Anatolia may have caused crustal thinning that generated the Cilicia-Adana trough.

The interpretation we propose for the evolution of the Carpathian and Aegean zones has some potential weak points. One is that not all authors recognize the decoupling fault system between the Anatolian and Iranian sectors of the Tethyan belt. This uncertainty could be due to the complexity and lateral width of such a transpressional shear zone. We think this shear zone may include the eastern sector of the north Anatolian fault system, which has become the main northern boundary for the westward escape of Anatolia since the early Pliocene. In this regard, we envision that a complex transpressional belt, extending from eastern Anatolia to Thrace, was active in the early Miocene (Perinçek, 1991; Andrieux et al., 1995; Sakinc et al., 1999). In the central Pontides, the present north Anatolian fault system runs within the southern part of the previously mentioned tectonic belt, whereas the northern part of the same belt became inactive (Andrieux et al., 1995).

Another difficulty for our interpretation could be the uncertainty that surrounds the onset of the Arabia's indentation, which some authors (e.g., Yılmaz et al., 1987; Yılmaz, 1993) place in the late Miocene–early Pliocene, much later than we suggest. This timing is based on the fact that the final stage of shortening with thrusting and uplift in the Bitlis suture zone occurred in the late Miocene-–early Pliocene, just preceding the formation of the north and east Anatolian fault systems, facilitating the westward extrusion of Anatolia (e.g., Şengör and Kidd, 1979; Koçyiğit et al., 2001). This interpretation implies that during much of the Miocene, the shortening across the Bitlis zone has accommodated most of the north-south convergence between Arabia and Anatolia (Yılmaz et al., 1987). However, this hypothesis cannot easily explain the early–middle Miocene shortening widely recognized north of the Bitlis zone, in the Pontides and Caucasus (e.g., Zonenshain et al., 1990; Andrieux et al., 1995; Kopp and Shcherba, 1998; Saintot and Angelier, 2002; Ershov et al., 2003; Nikishin et al., 2003).

In our opinion, this evidence implies that eastern Anatolia has moved westward with a significant northward component with respect to Eurasia since the early Miocene, and we interpret this mobility as an effect of Arabia's pushing after the continental collision along the Bitlis suture zone.

The occurrence of thrusting and uplift along the northern border of the Arabian indenter in the late Miocene–Pliocene (e.g., Şengör and Kidd, 1979; Yılmaz et al., 1987; Koçyiğit et al., 2001) can be interpreted as an effect of the complete consumption of the low-buoyancy lithosphere that lay north of Anatolia. The strong constrictional regime in front of the Arabian indenter, induced by the previously mentioned suture, has necessitated the consumption of continental lithosphere, emphasizing crustal thickening and uplift in the Bitlis zone and eastern Anatolia.

Further support for the hypothesis that stress can be transmitted from one continental block to another, once the intervening oceanic zone has been consumed, can be obtained from the most recent evolution of the Hellenic arc. Some authors recognize the effects of the incipient continental collision between the Hellenic arc (Crete) and the Libyan promontory of the Africa plate. In Figure 5 we describe the considerable tectonic reorganization of the Aegean tectonic setting that, in our opinion, was controlled by the previously mentioned incipient collision. However, in spite of the fact that such "interaction" has already produced evident effects in the southern Aegean zone since the late Pliocene (e.g., Armijo et al., 1992; Le Pichon et al., 1995; Piper and Perissoratis, 2003; ten Veen, 2004; Mantovani, 2005), at present the colliding blocks (Crete and Libya) are still separated by a marine zone, in which surface deformation involves only thrusting and folding of the Ionian-Levantine sedimentary cover subducting beneath the Hellenic arc (the so-called Mediterranean Ridge).

During the Miocene, the Africa-Adriatic continental domain and the Anatolian-Cycladic–Pelagonian Tethyan belt converged at the expense of the intervening low-buoyancy oceanic lithosphere, the northern Ionian (or pre-Apulian) zone (Mercier et al., 1987). Figure 4B shows the tectonic and kinematic settings that characterized the latest period of this phase (late Miocene), just before the suturing of the pre-Apulian consuming boundary (e.g., Mercier et al., 1987).

Late Miocene to Late Pliocene

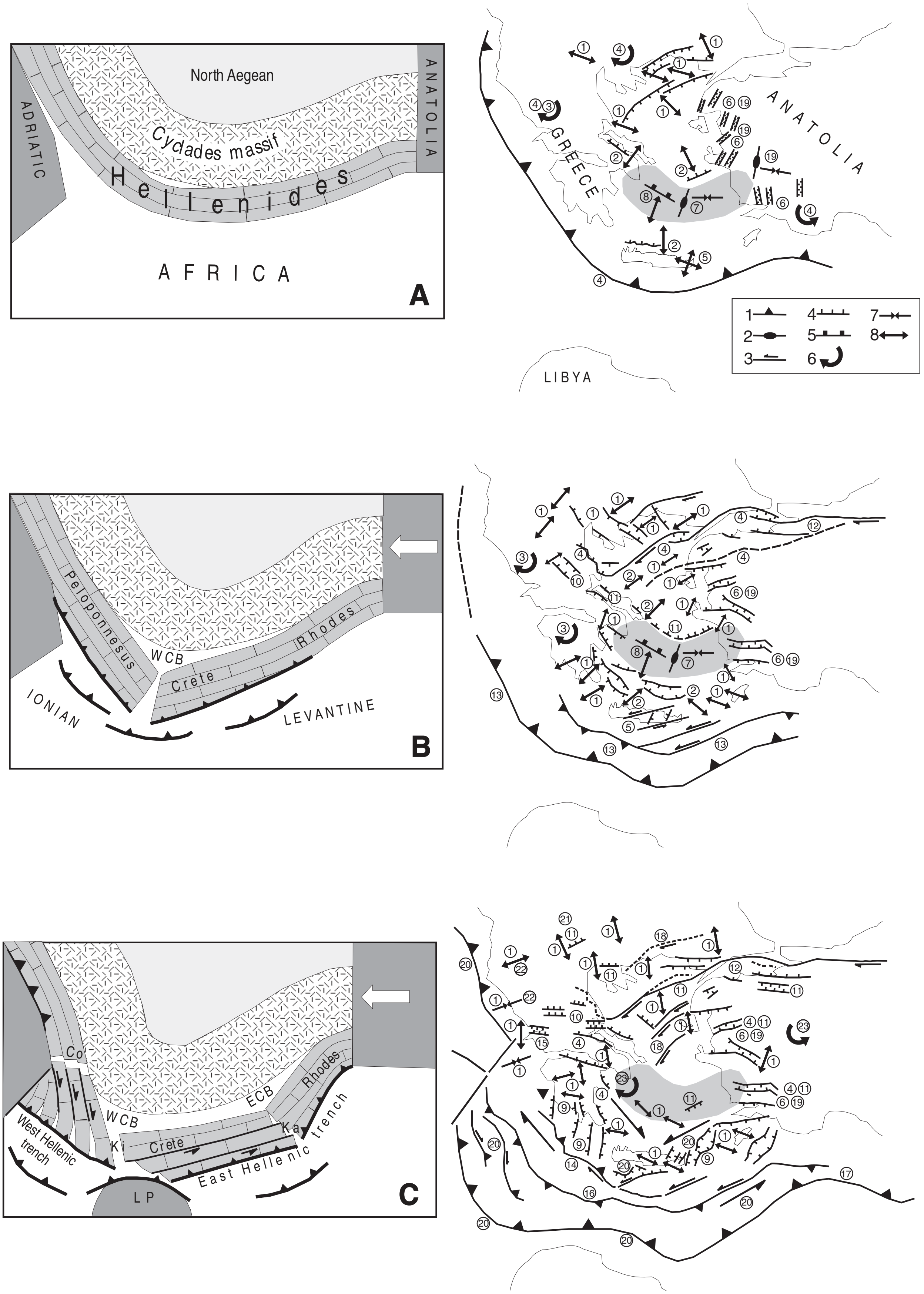
Around the Late Miocene, several major tectonic events occurred almost simultaneously in the central and eastern Mediterranean, causing a profound structural reorganization of the region (Fig. 6A).

Central Mediterranean. In the Apenninic belt, with particular regard to the southern arc, orogenic activity was resumed after a relatively long period of quiescence (e.g., Di Nocera et al., 1976; Ortolani, 1979; Castellarin and Vai, 1986; Patacca and Scandone, 1989). Crustal stretching ceased in the northwestern Tyrrhenian but developed in the central Tyrrhenian with the opening of the Magnaghi-Vavilov basin (e.g., Sartori, 1990; Sartori and Capozzi, 1998). The Pelagian zone in the northern African margin was affected by transtensional tectonics, which led to the formation of troughs in the Sicily Channel (e.g., Finetti and Del Ben, 1986; Reuther, 1987; Argnani, 1993). Thrusting took place in the Maghrebian belt lying north of the Iblean-Ventura block (Grasso et al., 1992; Catalano et al., 1994). Vertical movements reactivated, since the late Miocene, the Syracuse and Apulian escarpments, which are the transition zones from the Ionian oceanic lithosphere to the continental lithosphere of the Iblean plateau on one side and the Adria plate on the other side (Carbone et al., 1982; Finetti, 1982; Auroux et al., 1984; Finetti and Del Ben, 1986). A major transtensional fault system (the Medina and Victor Hensen faults; see Fig. 6A) developed in the central Ionian area since the late Messinian (Hieke and Wanninger, 1985; Della Vedova and Pellis, 1989; Hieke and Dehghani, 1999). A major left lateral shear zone, the Schio-Vicenza fault, formed in the northern Adriatic foreland and the central–eastern Alps, and the orientation of shortening in the eastern southern Alps changed from roughly northeastward to northwestward (e.g., Castellarin and Vai, 1986; Cantelli and Castellarin, 1994).

We argue that the previously described tectonic events may be explained coherently as effects of the suture of the pre-Apulian consuming boundary (Epirus) between the continental Adriatic domain and the Cycladic-Pelagonian sector of the Tethyan belt. Geological data suggest that this suture developed around the late Miocene, preceded by considerable crustal thickening and uplift (Doutsos et al., 1987; Mercier et al., 1987). When this suturing took place (Fig. 4B), the Adriatic continental domain faced, on its western side, a relatively extended low-buoyancy lithospheric domain, the western Apulian zone (e.g., Serri et al., 1993; Beccaluva et al., 1994; Francalanci and Manetti, 1994; Finetti et al., 1996, 2001). In such a context, one can expect that the consumption of the western Apulian oceanic zone was the most efficient way to accommodate the convergence of the confining plates (Africa, Adria, and Eurasia) once the Adriatic-Aegean border had sutured. This consumption actually occurred, as testified to by the large amount of accretionary material that accumulated at the Apenninic and Calabrian consuming boundaries during the Pliocene (e.g., Patacca et al., 1990; Ortolani et al., 1992). In our opinion, the key tectonic event that determined such consumption was the counterclockwise rotation of the Adria plate (Fig. 6A), driven by the roughly westward push of the Cycladic-Pelagonian Tethyan belt. This rotation started around the late Messinian, once development of major faults allowed the Adriatic block to decouple from its northwestern (Padanian) protuberance and from Africa. The first decoupling was accommodated by the sinistral Schio-Vicenza fault system (Fig. 6A). This is suggested by the onset of this fault in the late Miocene and by the fact that the orientation of the compressional axis in the eastern southern Alps changed from northeastward to northwestward (Cantelli and Castellarin, 1994). Decoupling of the Adria plate from Africa (Fig. 6A) was accommodated by the formation of a major sinistral fault zone, the Victor Hensen–Medina–Sicily Channel fault system (Hieke and Wanninger, 1985; Della Vedova and Pellis, 1989; Hieke and Dehghani, 1999), coeval with the Schio-Vicenza fault.

The counterclockwise rotation of the Adria plate induced a strong compressional regime in the central Mediterranean area that was accommodated by the lateral escape of the Iblei-Ventura microplate (a fragment of Africa) and other crustal wedges of the Apennines and Maghrebides belts at the expense of the western Apulia and Ionian low-buoyancy lithosphere. This hypothesis (Mantovani, 2005) provides a coherent and plausible explanation for the time-space distribution of deformation in the Tyrrhenian-Apennines region and its vicinity from the late Miocene to the late Pliocene, as follows:

- The northwestward expulsion of the Iblei-Ventura microplate, guided by two lateral strike-slip faults (Fig. 6A), could be an effect of the roughly east-west convergence between the southern Adriatic–northern Ionian block and the African continental domain (Tunisia). Along the southern guide of the previously mentioned extruding wedge, the


North Aegean
ADRIATIC
ANATOLIA
Cyclades massif
Hellenides
AFRICA
A
GREECE
ANATOLIA
LIBYA
1
2
3
4
5
6
7
8
Peloponnesus
WCB
Crete
Rhodes
IONIAN
LEVANTINE
B
Co
WCB
ECB
Ki
Ka
Crete
Rhodes
West Hellenic trench
East Hellenic trench
LP
C

transtensional Sicily Channel fault system, developed with the formation of pull-apart troughs along the releasing strands of this fault system (e.g., Finetti 1982; Finetti and Del Ben, 1986; Reuther, 1987; Argnani, 1993).

- In turn, the roughly northwestward indentation of the extruding Iblei-Ventura wedge into the Alpine-Apenninic belt lying east of Sardinia caused eastward escape of orogenic crustal wedges at the expense of the western Apulian and Ionian low-buoyancy lithosphere (Fig. 6A). In front of these wedges, orogenic activity developed in the Apenninic belt with particular regard to the southern Apennines and the Calabrian arc, and extensional tectonics occurred in the wake of the same wedges, causing the formation of the Magnaghi-Vavilov basin in the central Tyrrhenian. The timing of crustal stretching (from latest Miocene to early Pleistocene), the east-west tensional trend, and the peculiar shape of the Magnaghi-Vavilov basin (e.g., Sartori and Capozzi, 1998) are consistent with the previously mentioned interpretation.
- The reactivation, around the late Miocene, of relative vertical movements along the Syracuse and Apulian escarpments might have been an effect of the downward flexure and progressive retreat that the northern Ionian lithosphere underwent under the load of the extruding Calabrian wedge.

Figure 5. The proposed geodynamic evolution of the southern Aegean region and surroundings since the late Miocene. (A) Late Miocene: The east-west convergence between Anatolia and the Adriatic is accommodated by the southward bowing of the Cycladic and Hellenic arcs. Extensional tectonics develops in the region lying north of the Cycladic arc. (B) Middle Pliocene: In response to bowing, the external belt (Hellenides) breaks into two sectors: the Peloponnesus and the Crete-Rhodes sectors. The divergence between the Cycladic arc and the previously mentioned two sectors is accommodated by crustal thinning, which generates the western Cretan basin (WCB). (C) Pleistocene: The incipient collision of the Libyan promontory (LP) of Africa with Crete forces the Crete-Rhodes sector to bend or extrude southeastward, at the expense of the Levantine low-buoyancy zone. The divergence between the Crete-Rhodes arc and the Cyclades massif induces crustal thinning, which generates the eastern Cretan basin (ECB). Clockwise rotation of blocks and internal deformation of the Peloponnesus, due to its oblique collision with the southernmost Adriatic continental plate, causes the formation of the Corinth trough (Co). After the Libya-Crete collision, the divergence between the Peloponnesus, still moving southwestward, and Crete (slowed down by the collision) produces the formation of north-south-trending troughs, such as the Kithira trough (Ki). Ka—Karpathos trough. Tectonic maps on the right column show the major deformation recognized in the Aegean area during the respective evolutionary phases on the left. Key to numbering of symbols: (1) Thrust; (2) fold axis; (3) strike-slip fault; (4) normal fault; (5) low-angle detachment; (6) sense of rotation of tectonic blocks; (7, 8) shortening and lengthening principal axes of horizontal strain, deduced from meso-structural analyses. The shaded zone indicates the Cyclades massif. The numbers next to symbols correspond to the following references: (1) Mercier et al. (1989); (2) Mascle and Martin (1990); (3) Kissel et al. (1995); (4) Armijo et al. (1996); (5) ten Veen and Postma (1999); (6) Yılmaz et al. (2000); (7) Avigad et al. (2001); (8) Sánchez-Gómez et al. (2002); (9) Armijo et al. (1992); (10) Caputo and Pavlides (1993); (11) Hatzfeld (1999); (12) Rangin et al. (2004); (13) ten Veen (2004); (14) Truffert et al. (1993); (15) Le Pichon et al. (1995); (16) Kopf et al. (2003); (17) Garfunkel (1998); (18) Goldsworthy et al. (2002); (19) Bozkurt (2003); (20) Kreemer and Chamot-Rooke (2004); (21) Doutsos and Koukouvelas (1998); (22) Papazachos et al. (1999); (23) Westaway (1990).

We argue that the major tectonic events that have taken place in this region since the late Miocene (Fig. 6A) can be coherently explained as artifacts of the roughly east-west compressional regime between Anatolia and the Adria plate after suturing along the pre-Apulian consuming boundary, as suggested in the following:

- During the Pliocene, accretionary activity accelerated along the external front of the Aegean arc, leading to the formation of the Mediterranean Ridge (e.g., Finetti, 1976; Underhill, 1989; Mascle et al., 1999; Huguen et al., 2001; Kopf et al., 2003). This deformation is consistent with the acceleration of the outward migration of the Hellenic arc induced by the previously mentioned suture (Fig. 6A).
- Transtensional tectonics accelerated in the north Aegean and the western Anatolian regions around the late Miocene–Early Pliocene, leading to the formation of a system of northeast-southwest-oriented dextral strike-slip faults (Hempton, 1987; Mercier et al., 1989; Taymaz et al., 1991; Armijo et al., 1999; Burchfiel et al., 2000; Koukouvelas and Aydin, 2002; Rangin et al., 2004). This kind of deformation can be expected in a dynamic context simultaneously characterized by east-west compression, the one induced by the convergence between Anatolia and the Adriatic and by north-south extension produced by the divergence between the Aegean arc and the Rhodope (Fig. 6A), as already suggested by other authors (e.g., Şengör and Kidd, 1979; Şengör and Yılmaz, 1981).
- Crustal thinning with a dominant north-south extensional trend mainly occurred in the western Cretan basin from the late Miocene to the late Pliocene (e.g., Angelier et al., 1982; Lyon Caen et al., 1988; Mercier et al., 1989; Mascle and Martin, 1990; Meulenkamp et al., 1994). The fact that crustal stretching occurred only in such a limited zone, with an almost triangular shape (Fig. 6A), and developed only during this limited period, imposes strong constraints on the driving mechanism of this tectonic event. Figure 5 shows a dynamic/kinematic scheme that could explain these peculiar features. This scheme suggests that the western Cretan basin evolved as a result of the divergence between the inner massifs (Cyclades) and the external arc (Hellenides). While the bending of the Cyclades arc in response to an east-west compression did not involve any major interruption of its continuity, the Hellenides belt, possibly due to its stronger curvature (being located in the outer part of the arc) and/or to its higher fragility, broke into two sectors, the Peloponnesus and Crete-Rhodes, which consequently diverged from the inner Cyclades belt (Fig. 5B). This inter-

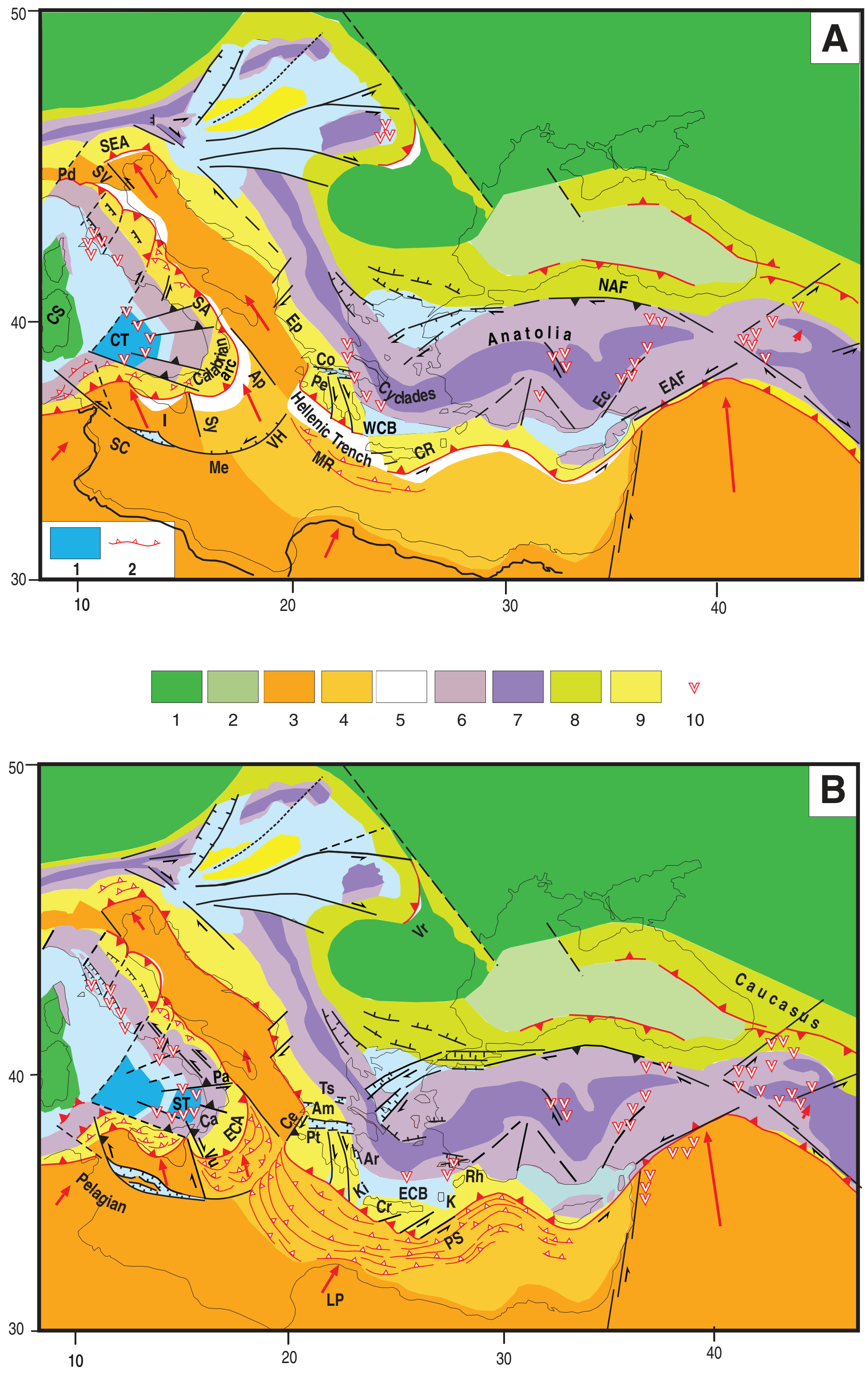


Figure 6. (A) The late Pliocene paleogeographic setting. The blue areas depict the zones affected by intense crustal thinning. Ap—Apulian escarpment; Ce—Cephalonia fault; Co—Corinth trough; CR—Crete-Rhodes sector of the Hellenic arc; CS—Corsica-Sardinia microplate; CT—central Tyrrhenian (Magnaghi-Vavilov basin); EAF—east Anatolian fault system; Ec—Ecemis fault; Ep—Epirus; I—Iblei-Ventura microplate; Me—Medina fault system; MG—Maghrebides belt; MR—Mediterranean Ridge; NAF—north Anatolian fault system; Pd—Padanian sector of the Adria plate; Pe—Peloponnesus; SA—southern Apennines; SC—Sicily Channel fault system; SEA—southeastern Alps; SV—Schio-Vicenza fault; Sy—Syracuse escarpment; VH—Victor-Hensen fault system; WCB—western Cretan basin. (B) The late Pleistocene paleogeographic setting. Am—Ambracique trough; Ar—Argolikos gulf; Ca—Calabria; Cr—Crete island; ECA—external Calabrian arc; ECB—east Cretan basin; K—Kirsehir massif; Ki—Kithira trough; LP—Libyan promontory; Pa—Palinuro fault; PS—Pliny and Strabo faults; Pt—Patras trough; Rh—Rhodes island; ST—southern Tyrrhenian (Marsili) basin; Ts—Thessaly trough; Vr—Vrancea zone; Vu—Vulcano fault. Other symbols and shading as in Figures 1, 2, 4, and 6A.

pretation provides a possible explanation for the timing of extension in the western Cretan basin and for the location and shape of that basin.

- The Cyclades massifs were affected by intense compressional deformation and uplift, which caused the first deposition of continental facies after the Miocene marine sedimentation (e.g., Angelier, 1977; Durr et al., 1978; Mercier et al., 1987, 1989; Boronkay and Doutsos, 1994). This scenario is consistent with the proposed east-west convergence between Anatolia and the Adriatic during this phase. Other possible effects of the east-west compression in the Cyclades zone, such as large-scale overturned and upright folds with axes parallel to the extensional trend (roughly south-north), have been recognized in the Cyclades islands (e.g., Robert, 1982; Urai et al., 1990; Buick 1991a,b; Bröcker et al., 1993; Ziv, 1996; Lister and Forster, 1997; Avigad et al., 2001; Rosenbaum et al., 2002). Furthermore, the fact that the development of crustal stretching for several million years, as suggested by the interpretation of exhumation processes (Gautier et al., 1999), has produced no net crustal thinning in this zone has been interpreted by Avigad et al. (2001) as an effect of contemporaneous thickening caused by the presumed east-west shortening.
- During this evolutionary phase, the inland segment of the Cephalonia fault behaved as a zone of transpressional dextral decoupling between the Peloponnesus and the Adria plate (e.g., Sorel, 1989; Mercier et al., 1989). The fact that the sense of shear (dextral) along this fault was opposite to that observed in the Victor-Hensen and Medina faults does not contradict the proposed kinematic interpretation (Fig. 6A), for deformation along this fault is not related to the relative motion between Africa and the Adriatic, but is rather an effect of the dextral relative motion between the Adriatic domain and the southwestward extruding western Hellenic arc.
- During this phase, the north Anatolian fault propagated westward up to the Aegean region (e.g., Barka, 1992; Westaway, 1994; Bozkurt, 2001; Armijo et al., 2004). This, along with the formation of the eastern Anatolian fault, determined the complete decoupling of the Anatolian wedge from the Eurasian domain and the start of its westward escape. The formation of the north and east Anatolian faults was preceded by thrusting, folding, and uplift in the collision zone where the high Anatolian-Iranian plateau formed (e.g., Şengör and Kidd, 1979; Yılmaz et al., 1987; Koçyiğit et al., 2001).

The previously mentioned deformation pattern and the extrusion of Anatolia could be understood as an effect of the minimum work principle. The complete consumption of low-buoyancy lithosphere in the Caucasus and in the pre-Apulian zone around the late Miocene considerably increased the resistance to any further northwestward displacement of eastern Anatolia. In this new context, the consumption of the low-buoyancy Ionian and Levantine domain became the main objective of tectonic reorganization, which was achieved by activating new fault systems (the northern and eastern Anatolian faults) for the westward escape of the Anatolian-Aegean wedge, leading to the present tectonic setting.

Quaternary

The beginning of a new evolutionary phase in both the central and the eastern Mediterranean regions around the late Pliocene–early Pleistocene (Fig. 6B) is indicated by the contemporaneous occurrence of several major tectonic processes.

Central Mediterranean. Accretionary activity came to an end in the southern Apennines and was accelerated in the Calabrian arc and the northern Apennines (e.g., Barone et al., 1982; Bigi et al., 1989; Patacca and Scandone, 1989; Sartori, 1989; Bartole, 1995). Extensional tectonics ceased in the central Tyrrhenian basin and began, with a northwest-southeast trend, in the southernmost Tyrrhenian, the Marsili basin (Kastens et al., 1988; Sartori, 1989; Sartori and Capozzi, 1998). Uplift accelerated in eastern Sicily, Calabria, and the southern Apennines (e.g., Ciaranfi et al., 1983; Westaway, 1993; Bordoni and Valensise, 1998). Tectonic activity accelerated in the Calabrian arc, leading to its strong fracturing and bowing (e.g., Selli et al., 1978; Ghisetti, 1979; Finetti and Del Ben, 1986; Van Dijk and Okkes, 1991; Del Ben, 1993; Meloni et al., 1997).

These major tectonic events can be explained as an artifact of suturing between the Adriatic continental domain and the extruding Apennines wedges along the southern Apennines' consuming boundary, documented by geological evidence (e.g., Casnedi et al., 1982; Patacca et al., 1993). No longer accommodated by this consuming process, the convergence of the confining plates caused the acceleration of the lateral escape of the Calabrian wedge, the only sector of the belt that still faced a low-buoyancy domain, the Ionian one (Fig. 6B). This extrusion was preceded and accompanied by crustal thickening and consequent uplift of the southern Apennines and the Calabrian arc. This interpretation can account for the deformation pattern in the Calabrian arc and surrounding zones since the late Pliocene, as follows:

- The cessation of crustal stretching in the central Tyrrhenian basin may be interpreted as a consequence of the halt of eastward extrusion of the southern Apenninic wedges.
- The acceleration of accretionary activity along the external front of the Calabrian wedge, with the formation of the external Calabrian arc, is consistent with the outward migration of the Calabria wedge.
- The timing of the extensional phase, which led to the formation of the Marsili basin, and the location and shape of this basin are compatible with the deformation expected in the wake of the escaping Calabrian wedge (Fig. 6B).
- The acceleration of tectonic activity at the Vulcano and Palinuro fault systems (e.g., Finetti and Del Ben, 1986; Finetti

et al., 1996), the two transcurrent faults of the Calabrian wedge, is consistent with the proposed extrusion process.
- The development of major transverse discontinuities and longitudinal troughs in Calabria and relative rotations of blocks may be seen as an effect of the southwest-northeast compression that drove this extrusion process (Fig. 6B).
- The acceleration of uplift that affected the zone directly involved in the Africa-Adriatic convergence, i.e., the Calabrian arc and southern Apennines, is compatible with the increase of compressional stress that followed the suture of the southern Apennines boundary.

The deformation pattern previously described—and, in particular, the acceleration of uplift in the Calabrian arc and the southern Apennines—have been explained as an effect of slab detachment under that zone (e.g., Cinque et al., 1993; Westaway, 1993). However, this interpretation is not consistent with the results of deep seismic surveys (Finetti, 2005), tomographic investigations (e.g., Piromallo and Morelli, 1997), and the study of high-frequency seismic wave propagation from deep Tyrrhenian earthquakes (Mele, 1998), all of which rule out any important interruption of the slab beneath the Calabrian arc.

Insights into the role played by the minimum work principle in controlling tectonic processes in the study area can be provided by the fact that accretionary activity was renewed at the southeastern Adriatic border (western Greece) around the early Pleistocene, after some million years of relative orogenic quiescence (e.g., Doutsos et al., 1987; Mercier et al., 1987). Such reactivation could indicate that lithospheric consumption at a plate boundary may stop in a given tectonic context, when it no longer represents the minimum work shortening process, but can be reactivated in a successive stage, when it has again become the most efficient process in the system. This can occur, for instance, when low-buoyancy lithosphere is considerably reduced or even absent in the surrounding zones. In this regard, it is useful to note that the suture of the pre-Apulian consuming boundary occurred in the late Miocene (after the complete consumption of the pre-Apulian low-buoyancy domain), when a large sector of low-buoyancy lithosphere (the western Apulian zone) was available on the other side of the Adria plate (Fig. 4). Then in the late Pliocene, after the complete consumption of this lithosphere (Fig. 6A), underthrusting of the Adriatic domain beneath the outer Hellenides could have again become the most efficient shortening process, in spite of the fact that continental Adriatic lithosphere was involved in the underthrusting.

Discussion on the compatibility between the recent evolutionary pattern proposed here and the major features of seismicity and recent volcanic activity in the central Mediterranean area has been reported by Mantovani et al. (1997b) and Tamburelli et al. (2000).

Eastern Mediterranean. Extensional tectonics ceased in the western Cretan basin and began, with a northwest-southeast to an east-west trend, farther to the east, leading to the formation of the eastern Cretan basin (Mercier et al., 1989; Armijo et al., 1992). The eastern segment of the Hellenic arc (Crete-Rhodes) underwent a southeastward migration and bowing (Duermeijer et al., 2000). The east-west to WNW-ESE extension and sinistral shear induced by this mechanism (Armijo et al., 1992) was probably responsible for the land interruption between Crete and Rhodes (e.g., Buttner and Kowalczyk, 1978; Meulenkamp et al., 1994). East-west to ENE-WSW extension and dextral shear have led to the formation of the Kithira and other north-south elongated troughs in the zone between the Peloponnesus and Crete (e.g., Lyberis et al., 1982; Armijo et al., 1992).

In our opinion, the key event which triggered the above tectonic reorganization was the arrival of a continental lithosphere at a subduction zone. This occurred around the Late Pliocene in the central Hellenic trench between Crete and the Libyan promontory of Africa, as suggested by geological and morphological evidence (Lyon-Caen et al., 1988; Armijo et al., 1992; Huguen et al., 2001; Piper and Perissoratis, 2003; ten Veen, 2004). After the beginning of that continental interaction, the southward motion of Crete slowed down considerably, emphasizing the rate of its convergence with Anatolia, which was accommodated by southeastward bending and extrusion of the Crete-Rhodes sector at the expense of the low-buoyancy Levantine zone (Figs. 5C and 6B). This mechanism, in analogy with what occurred in the Calabrian arc, also involved considerable fracturing and microblock rotation in the stressed sector, leading to the separation of Crete from Rhodes. The extensional tectonics that developed in between the extruding Crete-Rhodes arc and the inner Cyclades massifs led to the formation of the eastern Cretan basin (Fig. 5C). Another effect of the Crete-Libya collision could have been the generation of the ENE-WSW sinistral shear zone recognized between Crete and Rhodes (e.g., Mascle and Martin, 1990; ten Veen, 2004).

The east-west extension and dextral shear that have affected the zone between Crete and Peloponnesus since the late Pliocene could be a consequence of the fact that Crete, since its collision with the Lybian promontory, has become a sort of fixed hinge zone of the arc, whereas the Peloponnesus (still pushed by the Cyclades belt and still facing the Ionian low-buoyancy domain) has continued its soutwestward drift (Figs. 5C and 6B).

In the Peloponnesus, the east-west shortening induced by the convergence between the Tethyan belt and the Adriatic was accommodated by the southward escape of narrow crustal wedges at the expense of the Ionian low-buoyancy domain and by the relative rotation between Epirus and the Peloponnesus (Fig. 5C). This rotation produced the angular divergence and dextral shear that may be responsible for the formation of the Corinth, Patras, Ambracique, Argolis and Thessaly troughs (e.g., Berckhemer and Kowalczyk, 1978; Jackson et al., 1982; Doutsos and Poulimenos, 1992; Caputo and Pavlides, 1993; Armijo et al., 1996, 2004). The proposed mechanism (Fig. 5) can also explain the progressive westward propagation of the Corinth trough (e.g., Stiros, 1988; Doutsos and Piper, 1990). The hypothesis that this deformation was driven by plate convergence is corroborated by the fact that during the generation of the previously mentioned extensional features the Peloponnesus was affected by crustal thickening and uplift.

A peculiar aspect of the deformation pattern in the northwestern Greece arc is the simultaneous occurrence of compressional and tensional tectonics in a narrow zone (e.g., Doutsos et al., 1987; Papazachos and Kiratzi, 1996; Papazachos et al., 1999) thrusting along the coast of Epirus and of Albania and tensional features on the internal side of the Hellenides. This kind of behavior is similar to that observed in the northern Apennines arc (e.g., Elter et al., 1975), explained as an effect of bowing induced by belt-parallel compression on the basis of qualitative (e.g., Mantovani et al., 1997b, 2002) and quantitative (Viti et al., 2004) arguments. This kind of explanation could also be applied to the Hellenides belt, which is undergoing a belt-parallel compression induced by the westward push of Anatolia and transmitted by the Cyclades massifs (Fig. 5).

VOLCANISM AND UPPER CRUSTAL EXTENSION

Volcanic activity is one of the basic imprints of past tectonic processes. Thus, the timing of the volcanism and petrology of erupted magmas may provide information on the tectonic evolution of the zone involved. This evidence becomes even more important when subduction-related (SR) volcanic products are involved, for they constitute one of the few surface features that may help those identifying and studying deep tectonic processes. However, to avoid misleading interpretation of this evidence one should understand whether the distribution, in space and time, of subduction-related volcanism is related to deep tectonic processes or is rather controlled by the strain field in the upper crust. For instance, some authors (e.g., Lister, 1991; Rubin, 1995; Dahm, 2000; Tamburelli et al., 2000) have suggested, on the basis of field evidence, theoretical analysis, and numerical modeling of volcanic activity, that magmas can reach the surface only when and where extension in the upper crust generates major vertical fractures through which magmatic products can arise. This would imply that the eruption of subduction-related magmas might occur much later (on an order of millions of years) than the generation and accumulation of metasomatized magmas at depth.

These researchers argue that identifying a plausible and systematic causal connection between volcanism and lithospheric subduction, this last identified by the time development of accretionary activity at the Apennines consuming boundary, is not easy, and they point out a regular correspondence, in space and time, of such volcanism with crustal extension. The hypothesis that contamination of mantle magmas by subducted crustal material may have occurred much earlier than the SR volcanism in the Tyrrhenian-Apennines system has also been advanced by other authors (e.g., Coli et al., 1991).

Other insights into this problem may be inferred from geological and volcanological evidence in the eastern Mediterranean region. In eastern Anatolia, widespread late Miocene–Quaternary volcanic activity, evidenced by an extensive cover of calc-alkaline to alkaline rocks (Innocenti et al., 1982; Pearce et al., 1990; Keskin et al., 1998; Yılmaz et al., 1998), occurred well after the end of subduction at the Bitlis collision zone, which occurred in the late Eocene (Yılmaz, 1993; Robertson, 2000). Several authors have related this activity to extensional structures (tail cracks, horsetails, and releasing bends) associated with northwest-southeast dextral and northeast-southwest sinistral strike-slip faults (e.g., Yılmaz et al., 1987; Yılmaz, 1990; Adiyaman et al., 1998; Koçyiğit et al., 2001).

In central Anatolia, calc-alkaline volcanism has developed since the late Miocene (e.g., Pasquarè et al., 1988). This activity cannot easily be connected with a subduction process, for lithosphere subduction in this zone ceased earlier (e.g., Robertson, 2000). Several authors (Toprak and Göncüoglu, 1993; Deniel et al., 1998; Dhont et al., 1998; Jaffey et al., 2004) propose that the eruption of the previously mentioned SR volcanism was triggered by extensional tectonics along pull-apart basins or releasing bends that developed in this zone in response to the deformation and lateral migration of the Tethyan complex.

In western Anatolian and Aegean regions, volcanic activity with a calc-alkaline and shoshonitic character has developed since the early Miocene (e.g., Fytikas et al., 1984; Yılmaz, 1997; Aldanmaz et al., 2000; Pe-Piper, 2002). The location and timing of this activity have led the authors just mentioned to suggest that this phenomenon was associated with crustal extension. Our evolutionary reconstruction predicts that the zone where such volcanism took place was affected by coeval extension in response to the divergence between the southward-bowing inner Tethyan belt and the Rhodope (Fig. 4A).

Another example of the fact that the timing of volcanic activity may be controlled by extension in the upper crust rather than by the tectonics of the underlying slab may be the last phase of volcanism in the Aegean region that generated the present calc-alkaline volcanic arc located along the southern border of the Cyclades belt (Fytikas et al., 1984; Papadopoulos, 1989; Papazachos and Panagiotopoulos, 1993). This volcanism is coeval with the development of transtensional features in the southern Aegean since the early Pleistocene (Perissoratis, 1995; Piper and Perissoratis, 2003; Pe-Piper et al., 2005), possibly in response to the Crete-Libya collision. On the other hand, finding a causal relationship between this volcanism and the Hellenic subduction process that began at least 5 or possibly 13 m.y. earlier (Angelier et al., 1982) would be rather problematic.

In the Carpatho-Pannonian region, geochemical evidence suggests that a large volume of subduction-related magmas was stored above the subducting slab during the consumption of the Magura domain (Seghedi et al., 2004). The eruption of this material with a calc-alkaline (andesitic and silicic) character has occurred mainly in the back-arc zone (the Pannonian and Transylvanian basins) since the early Miocene (e.g., Csontos, 1995). It is worth noting that such activity was coeval with the extension that affected that zone in response to the differential displacement and relative rotation of the Alcapa and Tisza-Dacia blocks (e.g., Csontos and Nagymarosy, 1998; Seghedi et al., 2004). The most recent (9–1 Ma) volcanic phase in the eastern part of the Transylvanian basin has been interpreted as an effect of extensional tectonics in the wake of the southeastward-extruding Vrancea block, facilitated by major strike-slip faults (Hippolyte et al., 1999; Seghedi et al., 2004).

DISCUSSION

It is well known that any geodynamic interpretation can be only tentative, for the complexity of the problem and the only partial knowledge of several basic elements, such as past and present structural and tectonic features, does not allow a definitive demonstration of the proposed model. However, the elaboration of tentative models is an unavoidable preliminary step toward a satisfactory understanding of Mediterranean geodynamics. This tentative elaboration must be then followed by quantitative checking of the compatibility of the proposed geodynamics with the observed deformation pattern.

In this section, we report some considerations about the plausibility of the two most popular alternative driving mechanisms so far proposed for the study area and about possible drawbacks of the model proposed here. We also make some remarks about the ambiguity that could surround a crucial feature of the Oligocene paleogeographic configuration we have adopted (Fig. 2), i.e., the structure of the Tethyan belt.

Slab-Pull Model

Attempts at explaining the evolution of the Aegean trench arc–back-arc system as an effect of a slab-pull mechanism (e.g., Le Pichon and Angelier, 1979; Meulenkamp et al., 1988) have assumed that the observed deformation has been driven by the gravitational sinking of the Ionian-Levantine slab. The plausibility of this hypothesis has been investigated quantitatively by assuming rather different configurations of the driving force. For instance, Meijer and Wortel (1996, 1997) impose an arc-normal pull force to the whole Hellenic arc, whereas Cianetti et al. (2001) apply such a force to the western sector of the arc only. These different choices suggest that considerable uncertainty still persists about the geometry and magnitude of the presumed force and, in particular, of its horizontal component (trench suction), which is presumed to be responsible for slab roll-back.

Furthermore, these attempts neglect a major problem, the fact that the extensional stress induced in the overriding plate by trench suction is much lower than the average strength of the lithosphere (e.g., Shemenda, 1993; Becker et al., 1999). In any case, it is generally acknowledged that relatively short slabs (<300 km) cannot induce extension in the overriding plate, even when it is mechanically weakened (e.g., Hassani et al., 1997). Thus, the feasibility of the slab-pull model crucially depends on whether a well-developed lithospheric body exists beneath the Aegean region. The fact that no earthquakes deeper than 150–180 km have been recorded in the Benioff zone (e.g., Papazachos et al., 2000) would exclude such a possibility. However, tomographic investigations (e.g., Spakman et al., 1993; Bijwaard et al., 1998) suggest that a long slab, reaching a depth of 1000 km or even more, is present beneath the Aegean area. This implies that most of the retreating slab is aseismic, which would be a quite peculiar feature with respect to the known subduction systems (e.g., Kirby et al., 1996). Furthermore, if the onset of subduction is placed at 5 or 13 Ma, as suggested by some authors (e.g., Le Pichon and Angelier, 1979) and a reasonable plate convergence of 1–2 cm/y is taken into account in the Hellenic trench, one would expect a slab length not exceeding 100–200 km. On the other hand, it is not plausible to attribute the generation of the Hellenic slab to previous subduction processes, such as the one that led to the consumption of the northern Neotethyan Ocean, for the remnant subducted body would have a much greater lateral extension from eastern Anatolia to the Pelagonian zones, as indicated by the configuration of the Tethyan belt (Fig. 2). In that case, one should explain why slab pull acted only in the Aegean sector.

Important insights into this problem might be derived from tentative reconstructions of the paleo-oceanic domains that separated the African and Eurasian forelands in the Cretaceous (e.g., Şengör and Yılmaz, 1981; Dilek and Moores, 1990; Robertson et al., 1996; Dilek et al., 1999). These models, characterized by alternating continental and oceanic domains, suggest that the formation of a single long slab under the Aegean zone, such as the one suggested by Faccenna et al. (2003), would be very unlikely, because thick continental crust entering the trench zone can resist subduction (e.g., Gessner et al., 2001).

Other possible shortcomings of the slab-pull model are suggested by the fact that the strain pattern predicted by this driving force cannot easily be reconciled with the following observations: .

- Considerable crustal thickening and uplift that affected the outer Hellenides and Albanides during the recent evolution (Sulstarova et al., 1980; Doutsos et al., 1987) is not consistent with the implications of the slab-pull model. One should also explain why this zone, where no significant compressional deformation is predicted by such interpretation, is affected by very strong and frequent earthquakes with transpressional source geometries (e.g., Baker et al., 1997; Louvari et al., 1999; Viti et al., 2001).
- No important changes concerning the direction and rate of subduction are recognized in the consuming process that has taken place beneath the Hellenic arc since the late Miocene (e.g., Le Pichon and Angelier, 1979). Thus, it is not easy to explain why the temporal and spatial distribution of extension in the Aegean region was so heterogeneous in space and time (e.g., Mascle and Martin, 1990). Crustal thinning first occurred in a small zone (the western Cretan basin) with a roughly south-north trend during the Pliocene and has developed in the eastern Cretan basin only since the early Pleistocene, with a roughly northwest-southeast extensional trend (Mercier et al., 1989). One should

also explain why in this last period roughly east-west extension has developed in the zone between Peloponnesus and Crete and between Crete and Rhodes (e.g., Armijo et al., 1992), in spite of the fact that the slab-pull model would predict a southwest-northeast extensional trend, and why the Aegean basins (the Cretan Sea and north Aegean trough) are separated by relatively wide structural domains that have undergone minor or no thinning, such as the Cyclades massifs (e.g., Avigad et al., 2001; Li et al., 2003).
- Most recent modeling attempts acknowledge that trench suction alone is not sufficient to explain the complex recent or present strain pattern recognized in the Anatolia-Aegean region. For instance, the transtensional tectonics that affected the northern Aegean (e.g., Pavlides et al., 1990; Koukouvelas and Aydin, 2002) cannot be explained without taking into account the westward extrusion of Anatolia (Armijo et al., 2004; Flerit et al., 2004). Similarly, the sinistral shear recognized in the Strabo and Pliny fault systems in the eastern Hellenic arc cannot easily be reproduced as effects of slab-pull forces (ten Veen and Mejier, 1998; ten Veen and Postma, 1999).

Other arguments in support of the hypothesis that the extrusion mechanism can account for the observed features of trench arc–back-arc migrating systems better than can subduction-related models are provided by Mantovani et al. (2001a), who have extended the comparative analysis of the implications of the previously mentioned geodynamic models to the central–western Mediterranean and circum-Pacific regions.

Gravitational Spreading Model

This model assumes that the extension in the Aegean zone was induced by the strong contrast in crustal thickness between the Tethyan orogenic belt, presumably thermally weakened, and the southern Neotethys oceanic domain, the Ionian-Levantine zone (e.g., Seyitoğlu and Scott, 1992; 1996; Gautier et al., 1999; Jolivet, 2001). In our opinion, the implications of this model, quantified by laboratory experiments (e.g., Hatzfeld et al., 1997; Gautier et al., 1999), cannot easily be reconciled with the observed features, for the following reasons:

- The bulk extension induced by gravitational spreading does not allow for the considerable east-west shortening, crustal thickening, and very strong seismic activity recorded in northwestern Greece and Albania (Doutsos et al., 1987; Baker et al., 1997; Papazachos et al., 1999). Because this major evidence cannot simply be neglected, another simultaneous driving mechanism must necessarily be identified.
- Avigad et al. (2001) suggest that the NNE-SSW brittle extension recognized in the central Aegean (Cyclades) has been accompanied by roughly WNW-ESE shortening, accommodated by folding of the ductile lower crust. This hypothesis is based on a large set of field data (see the section headed “Late Miocene to Late Pliocene”) and postulates a mechanism that has successfully been applied to similar tectonic contexts (Mancktelow and Pavlis, 1994). The presence of compressional stresses more or less perpendicular to the crustal extension cannot easily be reconciled with gravitational spreading.
- Another bit of evidence that cannot easily be accounted for by this driving mechanism is the contemporaneous presence of compressional (along the coast) and tensional deformation within a very narrow belt in Albania and Epirus (Doutsos et al., 1987; Papazachos and Kiratzi, 1996; Papazachos et al., 1999).
- This model predicts that extension has mainly affected the thick Tethyan belt, whose exhumed remnants form the present Cyclades belt (Ring and Layer, 2003). Despite the very long duration of spreading (more than 20 m.y.) and the high extension rate (12–22 mm/yr) suggested by Gautier et al. (1999), the crust of the Cyclades is presently thicker than that of the surrounding domains (e.g., Papazachos and Nolet, 1997; Li et al., 2003). Moreover, no deep basin has been formed in the present central Aegean Sea, whose average depth does not exceed 200 m. On the contrary, the northern and southern Aegean basins, which were formed later, are considerably deeper than the Cyclades Sea, reaching depths of 1600 m and 2400 m, respectively (Mascle and Martin, 1990). The possibility that minor crustal thinning has affected the crust of the central Aegean since the middle Miocene is also supported by the absence of block tilting in the Cyclades islands (Avigad et al., 1998) and by the amount of horizontal stretching estimated for this zone (Li et al., 2003).
- The evolution of the external border of the spreading body predicted by modeling experiments (e.g., Hatzfeld et al., 1997; Gautier et al., 1999) is not consistent with the present shape of the Hellenic arc, constituted by a northwest-southeast trench zone, and a perpendicular strike-slip fault system (Fig. 5) cannot easily account for the complex distribution in space and time of crustal thinning (Fig. 5) involving the formation of the western Cretan basin in the Pliocene and, in the Quaternary, the opening of the eastern Cretan basin and the Kithira and Karpathos troughs and the generation of the ENE-WSW sinistral shear fault system in the southeastern Aegean sea (Mascle and Martin, 1990; Piper and Perissoratis, 2003; ten Veen, 2004).

It is not easy to understand why crustal spreading did not develop in Anatolia, in spite of the fact that such a zone was characterized by crustal thickening at least comparable to that of the Aegean zone and that it also faced the same oceanic lithosphere, that of the Levantine domain. Whitney et al. (2001) suggest that ~20 km of the upper crust has been removed during exhumation of the crystalline massifs in central Anatolia. Because the present crustal thickness in this zone is 35–40 km (e.g., Seber et al., 2001), it is plausible that before exhumation the crust was as thick as that in the Aegean zone at the presumed onset of spreading.

Possible Drawbacks of the Extrusion Model

Objections to the feasibility of the extrusion model in the study area have been raised by some authors. Gueguen et al. (1997) and Jolivet et al. (1998) have argued that such tectonic mechanism cannot be applied to the Tyrrhenian-Apennines system because the rate of migration of the Apenninic arc was greater than the rate of convergence between Africa and Eurasia. However, this argument is based on the arbitrary assumption that the previously mentioned plate convergence was the only driving mechanism, while neglecting other major forces, such as the westward push of the Anatolian-Cycladic-Pelagonian Tethyan belt.

Furthermore, the previously cited objection does not take into account that in our interpretation (Fig. 6) the retreat of the subduction zone was produced by the combination of two driving mechanisms: the westward motion of the Adria plate (pushed by the Anatolian-Aegean system) and the eastward escape of crustal wedges in the Apennines.

Other arguments against the applicability of the extrusion model to the Aegean-Hellenic system are based on space geodetic data (e.g., McClusky et al., 2000). These observations indicate that the Aegean zone is moving faster (roughly 30–40 mm/yr) than the Anatolian wedge (15–25 mm/yr), which would contrast with the hypothesis that the deformation pattern of the Aegean area is driven by the westward extrusion of Anatolia. However, it must be pointed out that the present-day velocity field indicated by geodetic data involves rates considerably greater than the middle- to long-term ones deduced by geological and seismic data. Measurements of fault offsets along the north Anatolian fault since the early Pliocene (Barka, 1992, and references therein; Bozkurt, 2001; Hubert-Ferrari et al., 2002; Allen et al., 2004) suggest velocities ranging between 5 and 10 mm/yr. Comparable rates have been obtained by analysis of the long-term features of seismic activity along the north Anatolian fault (Barka, 1992). Similar estimates at the eastern Anatolian fault system (Cetin et al., 2003) suggest a slip rate of 11 mm/yr in the last 2.5 million years. Analysis of the seismic strain rate at the Hellenic trench related to the last century suggests a convergence rate of 4–15 mm/yr (e.g., Jackson, 1994).

A possible explanation of such a marked difference between short- and long-term kinematics is provided by quantification of postseismic relaxation in the Anatolian-Aegean system induced by the sequence of very strong earthquakes that have occurred along the north Anatolian fault since 1939 (Mantovani et al., 2001c; Cenni et al., 2002). The possibility that short-term and long-term kinematics may be significantly different is supported by other theoretical arguments and observational evidence for various seismic zones of the world (e.g., Anderson, 1975; Rydelek and Sacks, 1990; Pollitz et al., 1998).

As mentioned earlier, some authors (e.g., Gautier et al., 1999; Jolivet, 2001) have argued that extension in the Aegean region cannot be interpreted as an effect of the Anatolia extrusion, because the tectonic extension is presumed to have started earlier (early Miocene) than the extrusion (upper Miocene). This hypothesis is mainly based on the assumption that the effect of Arabia's indentation on the deformation of the Tethyan and Eurasian zones facing this promontory began only in the late Miocene–early Pliocene, when the thrusting, folding, and uplifting took place in the collision zone and Anatolia was expulsed westward. However, we explain the occurrence of extensional tectonics in the northern Aegean region during the early Miocene (e.g., Burchfiel et al., 2000) as a result of the divergence between the Pelagonian-Cycladic arc and the Balkanides-Rhodope zone that was induced by the southwestward bowing of the Tethyan belt (Fig. 4).

Exhumation of Metamorphic Massifs in the Tethyan Belt

In the first phase of our reconstruction (Fig. 2), we assumed that the inner metamorphic core of the Tethyan belt was already exhumed. This assumption is based on the hypothesis that this orogenic belt was created in a compressional, syncollisional tectonic environment during the closure of the northern Neotethyan Ocean (e.g., Avigad et al., 1997; Whitney et al., 2001; Ring and Layer, 2003). This view is not necessarily shared by other authors (e.g., Lister et al., 1984; Gautier et al., 1999), who suggest that the high-pressure metamorphic rocks recognized in the Aegean zone were exhumed by crustal extension much later (since the early Miocene) than proposed in our model. However, several pieces of evidence point against this last hypothesis. The construction of pressure-temperature-time paths for blueschist and eclogite-facies rocks (e.g., Avigad et al., 1997; Trotet et al., 2001; Ring and Layer, 2003) suggests that exhumation of the Cyclades massifs was mostly related to syncollisional processes, whereas the postcollisional detachments played a minor role, being mainly restricted to the upper brittle crust (Rosenbaum et al., 2002). Various mechanisms, including thrusting and accretion from below (Katzir et al., 2000), vertical extrusion of wedges (Ring and Layer, 2003; Xypolias et al., 2003), and buoyancy-driven uplift (Okay et al., 1998; Ricou et al., 1998; Schmädicke and Will, 2003), may account for exhumation in the convergent setting preceding the Aegean extension.

In particular, the last of the previously mentioned mechanisms seems to be particularly suitable to explain the main features of exhumation processes. Laboratory experiments (Chemenda et al., 1995; Boutelier et al., 2004) suggest that the continental crust entering into a subduction zone can be dragged to depths of ~120 km, where it undergoes high-pressure metamorphism. Then buoyancy forces induce the uplift of subducted crustal slices, which move upward in the space between the converging plates. It is worth noting that uplifting of these slices occurs along shear zones whose geometry and kinematics are very similar to those of the crustal detachments recognized in many metamorphic core complexes, as documented from high-pressure metamorphic rocks exposed in the Himalayas, the Urals, and Oman (Boutelier et al., 2004, and references therein).

Regarding the timing of exhumation, it has been suggested that the high-pressure rocks of the Cyclades have been located at a shallow crustal level, or even at the surface (Syfnos island), since at least 30 Ma (Parra et al., 2002; Schmädicke and Will, 2003). In the Olympos-Ossa massif in Thessaly, syncollisional exhumation has occurred since 50 Ma (Lips et al., 2000). In western Anatolia, the Menderes Massif reached shallow crustal depths at ca. 35 Ma (Rimmelé et al. 2003), whereas the Kazdag massif located farther to the north was mostly exhumed at ca. 24 Ma (Okay and Satir, 2000). Moreover, the unconformable deposition of fluvial and lacustrine sediments over the metamorphic rocks of the Menderes massif (e.g., Yılmaz et al., 2000) strongly suggests that the exhumation of that complex may have occurred earlier than the early Miocene.

In northwestern and central Anatolia, the slow exhumation of metamorphic complexes has been interpreted as a syncollisional event in the absence of core complex detachments (e.g., Okay et al., 1998; Whitney et al., 2001), and unroofing of these massifs was supposedly achieved much earlier than in the late Oligocene. Cooling ages of 50–45, 47–40, and 35–32 Ma are estimated for the Aksaray, Kirşehir, and Akdağ massifs in central Turkey, respectively (Whitney et al., 2001). The exhumation of the Niğde massif farther south, which occurred considerably later (12–9 Ma), has been interpreted to be associated with strike-slip tectonics along the left lateral Ecemis fault (Figs. 4B and 6A) of the central Anatolian fault zone (Fayon et al., 2001).

CONCLUSIONS

We argue that the indentation of Arabia has caused considerable deformation in a large area, encompassing the Anatolian, Aegean, Greek, Balkan, Carpathian, and even central Mediterranean regions. Such a broad and far-reaching effect has been favored by the high crustal strength and the geometry of the Tethyan orogenic belt. The strong inner core of the Tethyan system (the Anatolian-Cycladic-Pelagonian metamorphic belt) played a critical role, allowing the transmission of collision-induced stresses for long distances.

The strength and cohesion of this belt is also underlined by the fact that the westward propagation of the North Anatolian fault system, after having broken more than a thousand kilometers of the Pontides accretionary belt, has been interrupted in the northern Aegean, where it intersected the Pelagonian sector of the Tethyan belt (Ginzburg et al., 1987).

The lateral displacement of Anatolia away from the Arabia-Eurasia collision zone was accommodated by distortion of the Tethyan belt, in terms of outward bowing and extrusion at the expense of low-buoyancy lithosphere. From the Oligocene to the middle Miocene, this process developed in the Carpathian arc at the expense of the Magura low-buoyancy domain and in the Aegean arc at the expense of the Ionian-Levantine low-buoyancy oceanic domain (Fig. 4A and B). These extrusion processes led to the development of the Carpathian-Pannonian trench arc–back-arc system and to the start of the same process in the Aegean arc.

This deformation pattern lasted until the late Miocene, when the boundary conditions of the Tethyan belt underwent important changes in the Carpathian arc as a result of the collision of the extruding wedges with the continental Eurasian domain, and in the northwestern Aegean arc where the Tethyan belt collided with the continental Adriatic domain. This new tectonic regime led to the acceleration of extrusion processes in the Anatolian-Aegean sector of the Tethyan belt. This process was facilitated by the westward propagation of the north Anatolian fault into the Aegean zone, which caused the decoupling of the present Anatolian-Aegean wedge from the Eurasian domain and the start of its westward drift (Fig. 6A).

After suturing along the pre-Apulian boundary (northwestern Greece) in the late Miocene, the westward push of the Cycladic-Pelagonian Tethyan belt directly acted on the Adriatic block, forcing it to decouple from Africa to experience a clockwise rotation. This kinematic development induced a strong constrictional regime in the central Mediterranean region, which was accommodated by the lateral escape of an African fragment (the Iblei-Ventura microplate) and of Apenninic crustal wedges at the expense of the western Apulian and Ionian low-buoyancy lithosphere. These extrusion processes can account for the complex deformation patterns recognized in the central Mediterranean region since the late Miocene, in particular for the generation of the Tyrrhenian-Apennines trench arc–back-arc system.

Suturing along the pre-Apulian boundary also had important consequences for the deformation pattern of the Aegean region. In particular, the outward extrusion of the Hellenic arc was confined to the sector lying south of the southern limit of the Adriatic continental domain, the Cephalonia transpressional fault system.

Around the late Pliocene–early Pleistocene, tectonic activities changed considerably in the southern Aegean region as a result of the interaction between the Hellenic arc (Crete) and the Libyan promontory (Africa). In particular, this incipient continental collision accelerated the deformation of the Crete-Rhodes sector, which underwent southeastward bowing and extrusion. In the wake of this migrating arc, crustal stretching developed in the eastern Cretan basin.

Almost contemporaneously, an important change in deformation pattern occurred in the central Mediterranean region due to the collision between the southern Apennines and the continental Adriatic domain. Following this collisional event, the constrictional regime in that zone was mainly accommodated by the lateral escape of the Calabrian wedge, leading to the formation of the Marsili basin in the internal part of the arc and of the external Calabrian arc on the outer front of the migrating wedge.

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[AQ1]Our style sheet for the book calls for using a hyphen with “back-arc” but not “trench arc.”

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