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Collision tectonics of the Mediterranean region: Causes and consequences

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ABSTRACT

The late Mesozoic–early Tertiary evolution of the Mediterranean region was defined by a series of collisions between Gondwana-derived continental blocks and Eurasia as the intervening ocean basins closed. The late Tertiary–Quaternary evolution of the region was controlled by the generally northward motion of Afro-Arabia and the compressional tectonics induced by the convergence between Eurasia and Afro-Arabia. Earlier collisional events caused the formation of thick orogenic crust, high-standing plateaus, and heterogeneous mantle, and resulted in slab break-offs that were collectively crucial for the onset of postcollisional collapse of the mountain belts, tectonic extension, and magmatism. The diachronous collision of Adria (Apulia), as an appendage of Africa, with Europe along its irregular margins created the Alps, the Apennines, and the Dinaride-Albanide-Hellenide mountain belt at different times and affected the formation of the Carpathians in the east. The collision of the Arabian promontory with Eurasia ca. 13 Ma facilitated the westward tectonic escape of Anatolia and caused intense deformation taken up by crustal shortening and conjugate strike-slip fault systems in a zone of ~1000 km stretching from the Bitlis-Zagros suture zone in the south to the Greater Caucasus in the north. The Anatolia plate has been rotating counterclockwise relative to Eurasia during its escape to the west and hence has been experiencing internal deformation through a combination of strike-slip and normal faulting, including metamorphic core complex formation. Subduction roll-back along the Hellenic trench has likely been the driving force for this southwest motion of Anatolia and the extensional tectonics affecting the Aegean province in the upper plate throughout the late Tertiary. The widespread alkaline volcanism both in the Aegean extensional province and in the Turkish-Iranian plateau since the late Miocene shows chemical evidence for an enriched asthenospheric mantle melt source; in both regions postcollisional slab break-off events have played a major role in providing this asthenospheric material and weakening the orogenic crust significantly. Young basins (Tyrrhenian, Aegean?) are in the process of opening above strongly arcuate subduction zones in a broadly convergent system of the Mediterranean region. The mantle response to the discrete collisional events, the geometry of colliding continental margins, and the scale of collisions strongly controlled the syn- to postcollisional tectonics and magmatism in the Mediterranean region.

Keywords: continental collision, Mediterranean tectonics, slab break-off, lithospheric delamination, mantle flow, escape tectonics, postcollisional magmatism

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INTRODUCTION

During the late Mesozoic and the Cenozoic, the Mediterranean region experienced complex subduction events and associated collisions, resulting in mountain building and plateau formation, magmatism, foreland and hinterland deformation, foreland flexures and sedimentary basin evolution, escape tectonics, orogenic collapse and lithospheric-scale extension, and the opening of small ocean basins. The nature and duration of these events were strongly affected by the scale of collisions, the geometry of colliding continental margins, and the mantle dynamics and mantle response to subductions and collisions (Dewey et al., 1986; Jolivet et al., 1999; Spakman and Wortel, 2004). Continental collisions, preceded by ophiolite emplacement and the terminal closure of seaways, have involved the progressive impingement of buoyant continental blocks or plates with subduction zones and generally evolved in a diachronous fashion along the strike of the suture zones. In the Mediterranean region, most of the continental blocks that collided with Eurasia had a Gondwana affinity, and their tectonic history (rift, drift, collision, and orogeny) involved the evolution of Tethyan seaways and other interconnected basins between Eurasia and Gondwana and the relative motions between these two supercontinents (Dilek and Moores, 1990; Dilek et al., 1999a).

Studies of the seismic topography and shear wave splitting have been most helpful in imaging the mantle structure in and across the collision zones in the Mediterranean region. Tomographic studies have revealed the 3-D geometry of subducted slabs beneath the young collision zones and in modern convergent margins (Spakman et al., 1993; Wortel and Spakman, 2000; Spakman and Wortel, 2004), and the shear wave splitting analyses have helped us understand the role of mantle flow and the kinematic links between plate motions and the asthenospheric flow directions (e.g., Sandvol et al., 2003). Collectively, these subsurface observations made significant contributions to our reconstructions of the geodynamic evolution of the Mediterranean region and to our understanding of the crustal and mantle response to subduction and collision processes in ongoing orogenic zones.

This article presents a brief summary of the collisional events in the Mediterranean region in order to provide a spatial and temporal framework of the region for the other articles in this volume. It highlights the nature, timing, geometry, and crustal consequences of the major collisions that affected the geodynamic evolution of the region starting in the late Mesozoic.

COLLISION ZONES AND SUTURES

The Alps (Fig. 1 and Table 1)

This orogenic belt in south central Europe developed during two discrete collisional events. The first was the collision of Adria (or Apulia) with Eurasia in what is now known as the eastern Alps as the intervening small Mesozoic Meliata Ocean (Stampfli and Borel, 2004) closed during the Cretaceous (Table 1; Coward and Dietrich, 1989). This collisional event produced northwest-directed thrust sheets of the Austro-Alpine segment of the Alps (Fig. 2A). The second collisional event involved the collision of Adria with Eurasia in the Tertiary as the intervening Alpine Tethys closed, producing the west-northwest-vergent contractional structures and a strong imbrication of the Apulian (considered an appendage of Africa) and Eurasian crust in the western Alps (Coward and Dietrich, 1989; Dewey et al., 1989). This phase of the Alpine orogeny involved a significant amount of strike-slip deformation because of the counterclockwise rotation of Apulia (and Africa) with respect to Eurasia. Therefore, the Alpine orogeny evolved diachronously from east to west through time and involved nearly 250 km of European lithosphere in Alpine deformation.

Tomographic studies have revealed positive anomalies in the upper 150–200 km beneath the Alps and the northern Po Plain that are dipping to the south beneath the central Alps and to the east-southeast below the western Alps (Spakman and Wortel, 2004). These subsurface observations are consistent with the crustal architecture of collision-induced structures at the surface and point to a south-dipping European margin below the eastern Alps (Adria plate, Fig. 2A). The postcollisional Eocene–Oligocene magmatism (mostly granitic pluton emplacement) in the Alps has been interpreted to have resulted from detachment of the south-dipping Alpine Tethys lithosphere due to slab breakoff during and shortly after the collision in the western Alps (Davies and von Blanckenburg, 1995; Schmid et al., 1996).

The Apennines (Fig. 1 and Table 1)

This northwest-trending orogenic belt in the Apulian Peninsula of Italy is the youngest mountain system (Neogene-Quaternary) in the Mediterranean region, and its geometry and crustal architecture change significantly from north to south. It developed as a result of the collision of the western passive margin of Adria with the Apenninic platform and Europe along a west-dipping subduction zone (Doglioni et al., 1999; Lentini et al., this volume). The deformation front associated with this collision has migrated northeastward in the northern Apennines (Fig. 2B), eastward in the central southern Apennines (Fig. 2C), and southeastward in Calabria and Sicily (Fig. 2D), and the entire Apennines have experienced significant shortening (80–200 km) and uplift (up to 2.5 km) at the rates of ~1 mm/yr, and also considerable subsidence (up to 5 km) in the peri-Adriatic foredeeps (Cavazza et al., 2004, and references therein). At the southeastern end of the Apennines, the Ionian seafloor has been subducting beneath Calabria and Sicily, and the Tyrrhenian Sea has been developing as a Neogene backarc basin above this subduction zone (Fig. 2C and D). Thus, the Tyrrhenian Sea is a northwest-southeast-spreading young basin evolving in a broad continental collision zone. The Aeolian volcanoes along the

Figure 1. Simplified tectonic map of the Mediterranean region showing the plate boundaries, collisional zones, and directions of extension and tectonic transport. Red lines A through G show the approximate profile lines for the geological traverses depicted in Figure 2. MHSZ—mid-Hungarian shear zone; MP—Moesian platform; RM—Rhodope massif; IAESZ—Izmir-Ankara-Erzincan suture zone; IPS—Intra-Pontide suture zone; ITS—inner Tauride suture zone; NAFZ—north Anatolian fault zone; KB—Kirsehir block; EKP—Erzurum-Kars plateau; TIP—Turkish-Iranian plateau.

TABLE 1. MAIN COLLISION ZONES IN THE MEDITERRANEAN REGION AND THEIR TECTONIC AND GEODYNAMIC FEATURES

Orogenic belt	Plates or blocks in collision	Collision age (Ma)	Extensional deformation	Postcollisional magmatism	Geodynamic processes
Eastern Alps	Adria-Eurasia	Cretaceous	Insignificant	Granitic intrusions (Eocene–Oligocene)	Tectonic escape (dextral) along the Peri-Adriatic fault line
Western Alps	Adria-Eurasia	Tertiary	Cenozoic rift system of eastern and central Europe (Neogene–Pliocene); Rhone-Rhine graben system to the west and north	Granitic intrusions (Eocene–Oligocene); alkaline volcanism within the European foreland (middle Eocene–early Oligocene)	Slab break-off (south-dipping Alpine Tethys lithosphere)
Apennines	Adria–Apenninic platform and Europe	Neogene–Quaternary	Progressive ~ east-west opening of the Algero-Provencal (ca. 34–16 Ma) and Tyrrhenian (ca. 14 Ma–present) back-arc basins; rifting of Corsica-Sardinia from Europe	Alkaline volcanism in Sardinia, Campania, the Aeolian Islands, Mount Vulture (Plio-Pleistocene and Quaternary)	Slab detachment (northwest-dipping Adria lithosphere) beneath the central–southern Apennines; subduction roll-back in the Ionian Sea
Carpathians	Alcapa and Tsia–East European platform	Early to middle Miocene	Development of the Pannonian basin as a result of WSW-ENE back-arc rifting in the Miocene	Andesitic volcanism in the Apuseni Mountains and the eastern Carpathians (ca. 16 Ma–present) and alkaline volcanism in the eastern Carpathians (Quaternary)	Slab roll-back to the east; slab detachment (WSW-dipping East European platform); intrabasin (Pannonian) faulting as in escape tectonics
Dinarides-Albanides-Hellenides	Adria-Eurasia (Serbo-Macedonian and Rhodope-Pelagonia blocks)	Eocene (ca. 55 Ma)	Opening of northwest-southeast- (in the north) and ~ east-west-oriented (in the south) trans-tensional basins; Rhodope massif exhumed as a meta-morphic core complex (47–35 Ma)	Low-K calc-alkaline (5–4 Ma) and shoshonitic volcanism (2.5–1.5 Ma) in northern Greece and Macedonia	Slab roll-back along the Hellenic trench west of the southern Balkan Peninsula; large-scale dextral oblique faulting (Miocene–present) in the hinterland of the oblique collision zone
Western Pontides	Sakarya-Rhodope, Strandja, and Pontide continental blocks	Late Cretaceous–Eocene	Western Black Sea rifting (~north-south) as a back-arc basin opening	Calc-alkaline to alkaline volcanism in the Rhodope (39–26 Ma) and Strandja massifs and in the Thrace basin (Oligo-Miocene and younger)	Possible slab detachment (north-dipping Sakarya lithosphere); reactivation of the Intra-Pontide suture zone along the western segment of the North Anatolian fault after ca. 7 Ma
Western Anatolia	Sakarya-Antolide and Tauride continental blocks (Pelagonia's extension to the east)	Paleocene–Eocene	Metamorphic core complex exhumation (Kazdag and Menderes) starting in the Oligo-Miocene; ~ north-south extension and development of northwest-southeast and east-west graben systems; opening of the Aegean Sea as a back-arc basin above the north-dipping Hellenic subduction zone	Eocene and Oligo-Miocene magmatism (medium to high-K calc-alkaline); mildly alkaline volcanism (14–12 Ma); alkaline volcanism since 12 Ma; southern Aegean volcanism as an island arc system above the Hellenic subduction zone	Eocene slab break-off (north-dipping Anatolide-Tauride lithosphere); lithospheric delamination ca. 14–12 Ma and asthenospheric upwelling; subduction roll-back of the African lithosphere at the Hellenic trench; west-southwest tectonic escape of Anatolia along the North Anatolian fault zone
Central Anatolia	Sakarya-Kirsehir continental blocks in the north and Kirsehir-Anatolide and Tauride blocks in the south	Eocene	Nigde core complex formation starting in the Oligo-Miocene; exhumation of the Akdag and Kirsehir metamorphic massifs; development of the Cappadocia volcanic province in a broad extensional system	Eocene–Oliogocene high-K shoshonitic volcanism around the Çankiri basin and north of the Kirsehir massif; calc-alkaline to alkaline magmatism in the Kirsehir massif and within the Ulukisla-Sivas sedimentary basins (latest Cretaceous–Tertiary)	Eocene slab break-off (north-dipping Kirsehir and Anatolide-Tauride lithosphere slabs); sinistral oblique faulting along the southeast margin of the Kirsehir block (syn- to postcollisional)
Eastern Anatolia	Arabia-Eurasia (recently assembled Anatolia)	Middle Miocene (ca. 13 Ma)	East-west extension and oblique-slip faulting in the Turkish-Iranian high plateau; exhumation of the Bitlis-Pütürge metamorphic massifs in the collision zone	Calc-alkaline to alkaline volcanism (late Miocene through Plio-Quaternary), more alkaline toward the south in time; intraplate fissure eruption in the Arabian foreland (Pleistocene Karacadag shield volcano)	Miocene (or younger) slab break-off (north-dipping Arabian and/or Bitlis-Pütürge lithosphere); lithospheric delamination beneath the Eastern Anatolian accretionary complex and asthenospheric upwelling; west- and east-directed crustal displacement along conjugate fault systems

eastern edge of the Tyrrhenian Sea represent the volcanic front of this west-northwest-dipping Ionian subduction zone (Fig. 2C; Lentini et al., this volume). The continental collision zone in the southern Apennines continues into northern Africa as part of the larger Apennine-Maghrebian orogenic belt, where postcollisional extension and magmatism have been documented to be associated with slab break-off (Aite and Gélard, 1997; Maury et al., 2000; Coulon et al., 2002).

Seismic tomographic studies have shown that the slab architecture changes dramatically from north to south along the Apennine mountain chain. A steeply west-dipping short north Apennines slab has been imaged in the north, whereas the negative anomalies found at depths of 200 km beneath the central–southern Apennines have been explained by slab detachment within the west-southwest-dipping Adria lithosphere in the south (Fig. 2B and C; Spakman and Wortel, 2004). This slab detachment or tearing and associated asthenospheric upwelling has been responsible for recent and active volcanism in southern Italy (e.g., Mount Vesuvius, the Flegrini Field, and the Calabrian arc).

The Carpathians (Fig. 1 and Table 1)

The northwest-trending, arcuate-shaped Carpathian orogenic belt occurs at the eastern end of the Alps and represents an east-northeast-vergent fold and thrust system transported over the east European platform. To the south, the Carpathians are separated by the east-west-trending Balkanides along a complex wrench-fault system. Two major contractional deformation episodes affected the Carpathians: the latest Early Cretaceous thrusting of the inner Carpathians (made of Hercynian basement and Permian–Lower Cretaceous rock assemblages) and the late Oligocene–Miocene thrusting of the Outer Carpathians (made of Lower Cretaceous to lower Miocene turbiditic rocks) toward the foreland (Royden and Horváth, 1998; Fodor et al., 1999). The Pannonian basin west of the Carpathians has developed as a back-arc rifting above an eastward-retreating subducting slab along the Carpathian arc during the late early and middle Miocene (Fig. 2E; Nemcok et al., 1998; Fodor et al., 1999). The Carpathians and the Pannonian basin are divided into two segments along the northeast-trending mid-Hungarian shear zone (Fig. 1). The northward impingement of the Adriatic promontory in the latest Oligocene–early Miocene facilitated the separation of the northern and southern segments and the eastward extrusion of the northern segment along this shear zone, producing the arcuate shape of the Carpathians. The andesitic volcanism in the Carpathians and the Apuseni Mountains started around 16 Ma and migrated eastward in time, keeping pace with the slab rollback processes (Fig. 2E; Linzer, 1996; Seghedi et al., 2004).

The Dinarides, Albanides, and Hellenides (Fig. 1 and Table 1)

This northwest-trending orogenic belt occurs along the eastern edge of Adria (Apulia) and is connected with the southern Alps in the north. It displays a bidivergent structural architecture with east-vergent thrust fault systems in the internal zone to the east and west-vergent thrust faults in the external zone to the west (Fig. 2D). The east-directed thrusts are a result of ophiolite emplacement in the Latest Jurassic–Early Cretaceous, whereas the west-directed thrust faults are associated with the collision of Adria with Eurasia (Pelagonian platform) in the Tertiary (Dilek et al., 2005). The collision-induced deformation has been propagating westward into the Eocene–Quaternary sedimentary rock units in the Adriatic foredeep since the Eocene as in a thin-skinned fold and thrust belt development (Fig. 2D). Global positioning system measurements in this region indicate a north-northeasterly motion of Adria with respect to Eurasia that is oblique to the northwest-trending collisional front (Altıner et al., this volume). This oblique convergence between Adria and Eurasia along the Dinaride-Albanide-Hellenide orogenic belt has resulted in strain partitioning in and across a broad dextral shear zone in the Balkan Peninsula and facilitated the collapse of the orogenic belt in locally well-developed transtensional zones (Dilek and Koçiu, 2004; Dumurdzanov et al., 2005). The Dinaride-Albanide-Hellenide orogenic belt continues southeast into the Antalya-Lycian zone of the Taurides in southwest Turkey. Postcollisional magmatism is widespread in the Dinarides and the Hellenides. Low-K calc-alkaline volcanic rocks with ages of ca. 5.0–4.0 Ma and shoshonitic volcanics (mostly trachyte) with ages of ca. 2.5–1.5 Ma occur in the Voras Mountains in northern Greece and Macedonia (the former Yugoslavian Republic of Macedonia) (Pe-Piper and Piper, 2002).

The Balkanides (Fig. 1)

This east-west-trending, north-vergent thrust belt separates the Moesian platform to the north from the Rhodope massif to the south. It is characterized by extensive Mesozoic–early Tertiary flysch deposits, intense middle Eocene contractional structures, and relatively thick continental crust (~34–38 km) (Cavazza et al., 2004). The Rhodope massif represents a metamorphic core complex consisting of pre-Alpine and Alpine high-grade rocks intruded by Late Cretaceous to early Miocene granitoids and covered by Late Cretaceous to Tertiary volcanic and sedimentary sequences (Marchev et al., 2004 and this volume; Bonev, this volume). The southern part of the Balkanides (Srednogorie zone) includes a Late Cretaceous magmatic belt that extends from Romania and Bulgaria into Turkey (the Pontides) and the Lesser Caucasus and may represent a volcanic arc developed along the southern margin of Eurasia.

The Sutures and Collision Zones of Anatolia (Fig. 1 and Table 1)

The Anatolian Peninsula in the eastern Mediterranean region is a collage of continental blocks separated by ophiolites and suture zones, and these continental fragments, mostly derived from Gondwana to the south, were amalgamated through

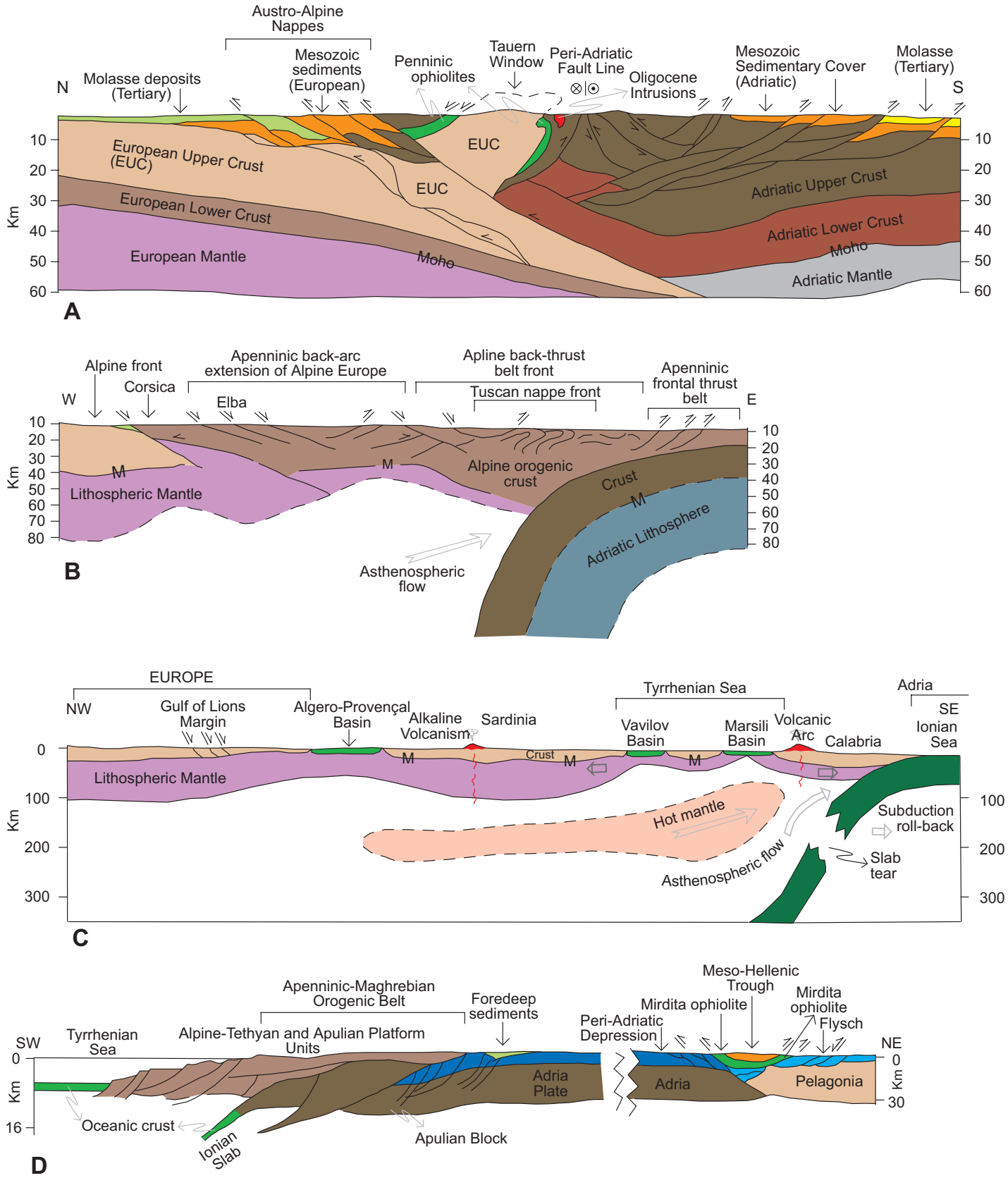
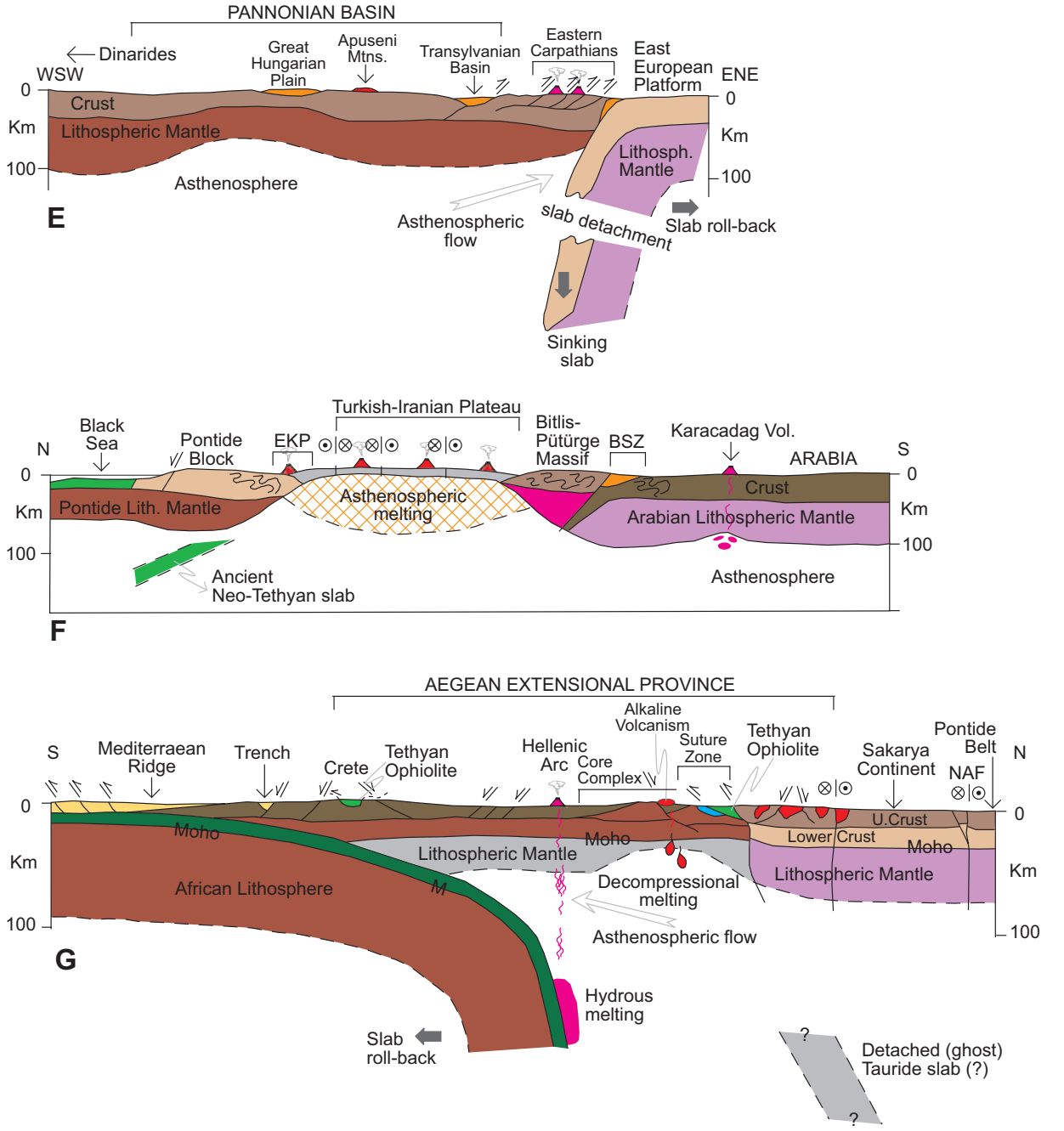


Figure 2. Simplified tectonic cross-sections across various segments of the broader Alpine orogenic belt. (A) Eastern Alps. The collision of Adria with Europe produced a bidivergent crustal architecture with both NNW- and SSE-directed nappe structures that involved Tertiary molasse deposits, with deep-seated thrust faults that exhumed lower crustal rocks. The Austro-Alpine units north of the Peri-Adriatic lineament represent the allochthonous outliers of the Adriatic upper crust tectonically resting on the underplating European crust. The Penninic ophiolites mark the remnants of the Mesozoic ocean basin (Meliata). The Oligocene granitoids between the Tauern window and the Peri-Adriatic lineament represent



the postcollisional intrusions in the eastern Alps. Modified from Castellarin et al. (2006), with additional data from Coward and Dietrich (1989); Lüschen et al. (2006); Ortner et al. (2006). (B) Northern Apennines. Following the collision of Adria with the Apenninic platform and Europe in the late Miocene, the westward subduction of the Adriatic lithosphere and the slab roll-back (eastward) produced a broad extensional regime in the west (Apenninic back-arc extension) affecting the Alpine orogenic crust, and also a frontal thrust belt to the east. Lithospheric-scale extension in this broad back-arc environment above the west-dipping Adria lithosphere resulted in the development of a large boudinage structure in the European (Alpine) lithosphere. Modified from Doglioni et al. (1999), with data from Spakman and Wortel (2004); Zeck (1999). (C) Western Mediterranean–Southern Apennines–Calabria. The westward subduction of the Ionian seafloor as part of Adria since ca. 23 Ma and the associated slab roll-back have induced eastward-progressing extension and lithospheric necking through time, producing a series of basins. Rifting of Sardinia from continental Europe developed the Gulf of Lion passive margin and the Algero-Provencal basin (ca. 15–10 Ma), then the Vavilov and Marsili sub-basins in the broader Tyrrhenian basin to the east (ca. 5 Ma to present). Eastward-migrating lithospheric-scale extension and necking and asthenospheric upwelling have produced locally well-developed alkaline volcanism (e.g., Sardinia). Slab tear or detachment in the Calabria segment of Adria, as imaged through seismic tomography (Spakman and Wortel, 2004), is probably responsible for asthenospheric upwelling and alkaline volcanism in southern Calabria and eastern Sicily (e.g., Mount Etna). Modified from Séranne (1999), with additional data from Spakman et al. (1993); Doglioni et al. (1999); Spakman and Wortel (2004); Lentini et al. (this volume). (D) Southern Apennines–Albanides–Hellenides. Note the break where the Adriatic Sea is located between the western and eastern sections along this traverse. The Adria plate and the remnant Ionian oceanic lithosphere underlie the Apenninic-Maghrebian orogenic belt. The Alpine-Tethyan and Apulian platform units are telescoped along ENE-vergent thrust faults. The Tyrrhenian Sea opened up in the latest Miocene as a back-arc (*continued*)

a series of collisional events starting in the Cretaceous (Şengör and Yılmaz, 1981; Dilek and Moores, 1990; Okay et al., 1996; Okay and Tüysüz, 1999; Ring et al., 2003; Kadioğlu et al., this volume).

The Intra-Pontide suture zone in northern Turkey is the result of a diachronous collision during the Cretaceous–Eocene between the Sakarya continent (to the south) and the Rhodope-Strandja continental block in the eastern Balkan Peninsula and between the Sakarya continent and the western Pontide belt in northwest Turkey. These continental entities collided as the intervening Intra-Pontide Ocean was consumed at a north-dipping subduction zone throughout the Cretaceous; this northward subduction resulted in the opening of the western Black Sea as a back-arc basin (Okay and Tüysüz, 1999). The Intra-Pontide suture has been reactivated by the North Anatolian fault zone (Fig. 1), which developed in the aftermath of the Arabian collision with Anatolia (Eurasia) around 13 Ma.

The Izmir-Ankara-Erzircan suture zone in central Anatolia has two segments. The western segment marks the collision zone between the Sakarya continent and the Anatolide-Tauride platform (the tectonic equivalent of Pelagonia in the Balkan Peninsula) that evolved along a north-dipping subduction zone within a Tethyan seaway. The southward-emplaced Cretaceous ophiolites and the Latest Cretaceous blueschist rocks along this suture zone constrain the polarity and timing of this Tethyan subduction zone (Okay et al., 1998; Sherlock et al., 1999). Following the ophiolite emplacement in the Latest Cretaceous, the continental collision occurred in the Paleocene–Eocene. Collision-induced southward thrust faulting was followed by the emplacement of postcollisional Eocene granitic plutons along the suture zone (Harris et al., 1994; Okay and Tüysüz, 1999; Altunkaynak and Dilek, this volume). The eastern segment of the Izmir-Ankara-Erzincan suture marks the collision zone between the Sakarya continent to the north and the Kirsehir and Anatolide-Tauride continental fragments to the south. The Paleozoic–Jurassic rocks of the Sakarya continental margin, Early Jurassic ophiolitic rocks, and Mesozoic seamount fragments are imbricated along generally south-directed thrust faults within this segment of the suture zone, forming a 30-km-wide belt of a colored mélange (i.e., the Ankara mélange) extending

Figure 2 continued basin behind the Apenninic-Maghrebian mountain belt. The Aeolian volcanoes in the Tyrrhenian Sea represent the volcanic arc system in this subduction-collision zone environment. Modified from Lentini et al. (this volume). The eastern section of this traverse across the Albanides-Hellenides in the northern Balkan Peninsula shows a bidivergent crustal architecture, with the Jurassic Tethyan ophiolites (Mirdita ophiolites in Albania and Western Hellenic ophiolites in Greece) forming the highest tectonic nappe, resting on the Cretaceous and younger flysch deposits of the Adria affinity to the west and the Pelagonia affinity to the east. Following the emplacement of the Mirdita-Hellenic ophiolites onto the Pelagonian ribbon continent in the Early Cretaceous, the Adria plate collided with Pelagonia-Europe obliquely starting around ca. 55 Ma. WSW-directed thrusting, developed as a result of this oblique collision, has been migrating westward into the peri-Adriatic depression. Modified from Dilek et al. (2005). (E) Dinarides–Pannonian basin–Carpathians. The Carpathians developed as a result of the diachronous collision of the Alcapa and Tsia lithospheric blocks, respectively, with the southern edge of the East European platform during the early to middle Miocene (Nemcok et al., 1998; Seghedi et al., 2004). The Pannonian basin evolved as a back-arc basin above the eastward-retreating European platform slab (Royden, 1988). Lithospheric-scale necking and boudinage development occurred synchronously with this extension and resulted in the isolation of continental fragments (e.g., the Apuseni mountains) within a broadly extensional Pannonian basin separating the Great Hungarian Plain and the Transylvanian subbasin. Steepening and tearing of the west-dipping slab may have caused asthenospheric flow and upwelling, decompressional melting, and alkaline volcanism (with an ocean island basalt–like mantle source) in the Eastern Carpathians. Modified from Royden (1988), with additional data from Linzer (1996); Nemcok et al. (1998); Doglioni et al. (1999); Seghedi et al. (2004). (F) Arabia-Eurasia collision zone and the Turkish-Iranian plateau. The collision of Arabia with Eurasia around 13 Ma resulted in (1) development of a thick orogenic crust via intracontinental convergence and shortening and a high plateau and (2) westward escape of a lithospheric block (the Anatolian microplate) away from the collision front. The Arabia plate and the Bitlis-Pütürge ribbon continent were probably amalgamated earlier (ca. the Eocene) via a separate collision event within the Neo-Tethyan realm. BSZ—Bitlis suture zone; EKP—Erzurum-Kars plateau. A slab break-off and the subsequent removal of the lithospheric mantle (lithospheric delamination) beneath the eastern Anatolian accretionary complex caused asthenospheric upwelling and extensive melting, leading to continental volcanism and regional uplift, which has contributed to the high mean elevation of the Turkish-Iranian plateau. The Eastern Turkey Seismic Experiment results have shown that the crustal thickness here is ~ 45–48 km and that the Turkish-Iranian plateau is devoid of mantle lithosphere. The collision-induced convergence has been accommodated by active diffuse north-south shortening and oblique-slip faults dispersing crustal blocks both to the west and the east. The late Miocene through Plio-Quaternary volcanism appears to have become more alkaline toward the south in time. The Pleistocene Karacadag shield volcano in the Arabian foreland represents a local fissure eruption associated with intraplate extension. Data from Pearce et al. (1990); Keskin (2003); Sandvol et al. (2003); Şengör et al. (2003). (G) Africa-Eurasia collision zone and the Aegean extensional province. The African lithosphere is subducting beneath Eurasia at the Hellenic trench. The Mediterranean Ridge represents a lithospheric block between the Africa and Eurasian plate (Hsü, 1995). The Aegean extensional province straddles the Anatolide-Tauride and Sakarya continental blocks, which collided in the Eocene. NAF—North Anatolian fault. South-transported Tethyan ophiolite nappes were derived from the suture zone between these two continental blocks. Postcollisional granitic intrusions (Eocone and Oligo-Miocene, shown in red) occur mainly north of the suture zone and at the southern edge of the Sakarya continent. Postcollisional volcanism during the Eocene–Quaternary appears to have migrated southward and to have changed from calc-alkaline to alkaline in composition through time. Lithospheric-scale necking, reminiscent of the Europe-Apennine-Adria collision system, and associated extension are also important processes beneath the Aegean and have resulted in the exhumation of core complexes, widespread upper crustal attenuation, and alkaline and mid-ocean ridge basalt volcanism. Slab steepening and slab roll-back appear to have been at work resulting in subduction zone magmatism along the Hellenic arc. Data from Jolivet and Patriat (1999); Aldanmaz et al. (2000); Pe-Piper and Piper (2002); Ring and Layer (2003); Ring et al. (2003); Altunkaynak and Dilek (this volume).

into the Lesser Caucasus in the east (Floyd, 1993; Tankut et al., 1998; Dilek and Thy, 2006). The Izmir-Ankara-Erzincan suture zone makes a sharp loop around the Eocene-Miocene Çankiri sedimentary basin (just east of the city of Ankara), which developed as a syn- to postcollisional foreland basin.

The inner Tauride suture in south central Anatolia marks the collision zone between the Kirsehir continent and the Anatolide-Tauride platform (Dilek et al., 1999b). The Late Cretaceous ophiolites rest tectonically on the Tauride platform along south-directed thrust faults, and there are discontinuous exposures of blueschist rocks exposed along the northern edge of the Tauride platform within this suture zone (Dilek and Whitney, 1997). The collision occurred in the Eocene, following the latest Cretaceous ophiolite emplacement at a north-dipping subduction zone, which consumed the inner Tauride ocean floor. This subduction zone was also responsible for the emplacement of the syncollisional granitic and syenitic plutons in the western part of the Kirsehir continent in the upper plate (Kadioğlu et al., this volume).

The Bitlis-Zagros suture defines the collision front between the Arabian and Eurasian plates that was initiated after the terminal closure of a Neo-Tethyan seaway, which had evolved between the Afro-Arabia and Pelagonia-Anatolide-Tauride platforms throughout the Mesozoic. The initial collision took place around 13 Ma, as constrained by local and regional structural, stratigraphic, and tectonic data (Şengör and Yılmaz, 1981; Dewey et al., 1986). Currently, Arabia is moving northwest relative to Eurasia at a rate of 20 ± 3 mm/yr to 24 ± 2 mm/yr (Reilinger et al., 1997). The Anatolia plate north of the suture zone has been displaced westward along the North and East Anatolian fault zones in the aftermath of the Arabian collision as part of the escape tectonics in the region (Jackson and McKenzie, 1984; Dewey et al., 1986). In its escape to the west, the Anatolia plate has been rotating counterclockwise relative to Eurasia, experiencing internal deformation (Piper et al., this volume) and overriding the eastern Mediterranean at two intermediate-depth seismic zones (the Cyprus and Hellenic subduction zones; LePichon and Angelier, 1979; Jackson and McKenzie, 1988; LePichon et al., 1995). Part of this counterclockwise rotation may be due to the differential motion of the Anatolia plate along the North and East Anatolian fault zones, with the rates around 30 ± 2 mm/yr and 15 ± 3 mm/yr, respectively (LePichon et al., 1995; Reilinger et al., 1997). However, the subduction pull and roll-back processes along the Hellenic trench are likely to be responsible for the southwest motion of the Anatolian plate toward the tectonic free face (Reilinger et al., 1997). As the Anatolia plate moves toward the Hellenic trench during its escape from the Arabian collision zone, it undergoes north-south lithospheric extension at 14 ± 5 mm/yr across the Aegean extensional province in the west (Fig. 1; Reilinger et al., 1997).

The hinterland of the Arabian collision zone is a high-standing plateau (the Turkish-Iranian plateau, Fig. 2F) with an average elevation of ~2 km above sea level, and is covered by upper Miocene–Quaternary calc-alkaline to alkaline volcanic rocks (Yılmaz et al., 1987; Pearce et al., 1990; Yılmaz, 1990; Keskin et al., this volume). The plateau displays structural evidence for active diffuse north-south shortening and east-west extension and a conjugate system of strike-slip faults (Fig. 2F) displacing crustal fragments toward Iran and the Caspian Sea (Jackson and McKenzie, 1984; Tan and Taymaz, this volume). The results of the recent Eastern Turkey Seismic Experiment across the east Anatolian plateau and the northernmost Arabia plate, combined with tomographic models of regional seismic velocity and attenuation, have shown that most of the plateau is lacking mantle lithosphere and that it is supported by hot asthenospheric mantle (Fig. 2F; Sandvol et al., 2003), not by thick crust as previously thought (Dewey et al., 1986). The absence of lithospheric mantle is interpreted to have resulted from break-off of northward-subducted slab beneath the east Anatolian accretionary prism. The extensive Pliocene–Quaternary volcanism in the region may be a consequence of melting of the lower crust above hot asthenosphere (Fig. 2F).

The collision-induced deformation farther north in the Erzurum-Kars plateau and the Caucasus region is taken up by both strike-slip and thrust faulting (Fig. 1). Lateral eastward displacements of crustal material along some major strike-slip fault systems (e.g., the Pompak-Sevan fault and the Van-Tebriz fault zone) have resulted in east-directed shortening and in the overriding of the continental crust of the Caucasus and northern Iran over the oceanic-type basement of the southern Caspian Sea (Jackson and McKenzie, 1984). The seismic patterns in the Caucasus indicate mainly shortening perpendicular to the northwest-southeast regional strike of the fold and thrust belt in this region and show that the collision-induced strain is partitioned across a nearly 1000-km-wide zone encompassing eastern Anatolia, northern Iran, and the Caucasus.

The prolonged collisional history of Anatolia, extending back to the latest Mesozoic–early Tertiary, played a significant role in crustal evolution and mantle dynamics in the eastern Mediterranean region during the late Tertiary (Dilek and Whitney, 2000). Specifically, the multiple episodes of continental collisions in western and central Anatolia produced thick orogenic crust and heterogeneous mantle that affected the mode and nature of postcollisional magmatism and tectonic extension in the region (Seyitoğlu and Scott, 1996; Yılmaz et al., 2001; Altunkaynak and Dilek, this volume). The first episode of postcollisional volcanism in western Anatolia evolved during the Eocene and the Oligocene–Miocene and produced medium- to high-K calc-alkaline rocks (Genç and Yılmaz, 1997; Altunkaynak and Dilek, this volume). This volcanic phase was also accompanied by widespread granitic plutonism in the region (Bingöl et al., 1982; Harris et al., 1994; Köprübasi and Aldanmaz, 2004) and was likely caused by an influx of asthenospheric heat and melts provided by slab break-off. The progressive change of the chemistry of volcanism from calc-alkaline to mildly alkaline to alkaline through time was accompanied by crustal- to lithospheric-scale extension and decreasing amounts

of subduction signature and crustal contamination. The main source of alkaline volcanism during the last 12–10 m.y. was decompressional melting of the asthenospheric mantle beneath the stretched continental lithosphere in the Aegean extensional province (Fig. 2G; Keller, 1983; Fytikas et al., 1984; Ercan et al., 1985; McKenzie and Yılmaz, 1991; Seyitoğlu et al., 1997; Savaşçin and Oyman, 1998; Aldanmaz et al., 2000; Alici et al., 2002). The onset of the Aegean subduction zone at the Hellenic trench around the middle Miocene (18–14 Ma; Meulenkamp et al., 1988) placed the entire Aegean region in a back-arc tectonic setting (Fig. 2G; LePichon and Angelier, 1979). Slab roll-back at the Hellenic trench is interpreted to have played a major role in the upper plate extension and exhumation of lower crustal rocks at the surface (Fig 2G; Jolivet et al., 1994; Lister and Forster, 1996; Barka and Reilinger, 1997; McClusky et al., 2000; Ring and Layer, 2003) and in subduction zone magmatism (Pe-Piper and Piper, 1989, 2002). The Kazdag and Menderes metamorphic core complexes in western Anatolia are the manifestations of postcollisional extensional tectonics in the eastern Mediterranean region (Fig. 2F; Hetzel and Reischmann, 1996; Bozkurt and Satir, 2000; Okay and Satir, 2000; Rimmelé et al., 2003; Gessner et al., 2004; Ring and Collins, 2005).

CONCLUDING REMARKS

The collisional tectonics of the Mediterranean region is highly complex and is mostly defined by the welding of continental fragments of all sizes and shapes into the Eurasian continental margin at different times during the late Mesozoic and the Tertiary. These continental fragments were detached from western Gondwana during the late Paleozoic–early Mesozoic and traveled northward (in the present coordinate system), facilitating the development of discrete ocean basins in their wake. The geological evolution of these ocean basins was influenced by the relative motions of the bounding supercontinents (i.e., Laurasia and Gondwana) with respect to each other, as well as by the motions of the individual blocks with respect to each other. The Tethyan systems (Paleo- and Neo-Tethys) had a far more complex paleogeography and geochronology than the previous geodynamic models envisioned (e.g., a single large Tethyan ocean versus two large Tethyan basins).

Continental collisions were preceded by ophiolite emplacement caused by impingement of trailing passive margins on the trenches and subduction zones. The majority of the eastern Mediterranean ophiolites were obducted onto the continental margins during the Cretaceous. The terminal closure of the basins and most of the continental collisions within the Tethyan domains appear to have occurred in a short time span during the early to middle Eocene, indicating that this was an important time window for tectonic clustering of continents and continental blocks in the eastern Mediterranean region.

Although the generally northward motion of Afro-Arabia caused the collapse of the Tethyan ocean basins, leading to continental collisions, some young basins (e.g., the Tyrrhenian and Balearic) developed within this broadly compressional system during the late Tertiary. Two factors seem to have played a major role in this development: (1) rapid roll-back of subducting lithosphere, causing upper plate extension, basin formation, and magmatism and (2) collisions between irregular continental margins, where strongly arcuate subduction zones spread to fill the gap. The young, thin, hot oceanic lithosphere in these modern basins constitutes the "provenance" of the future ophiolites in the broader Alpine-Himalayan orogenic system.

Slab detachment, slab tearing, and slab roll-back were all important components of subduction tectonics that were fundamental for the swift change from overall compression to extension in the Mediterranean collision zones. Slab detachment (break-off) was generally caused by partial subduction of buoyant continental edges and their isostatic rebound, while the colder, denser leading edge of the subducting oceanic lithosphere broke off and sank into the lower mantle. Slab tearing may have occurred along the intersection of strongly arcuate subduction zones (i.e., the Cyprean and Hellenic arcs), which may have had different convergence rates and/or different convergence directions. Both slab detachment and slab tearing were fundamental processes for facilitating slab roll-back, and collectively all three processes played a major role in asthenospheric upwelling and decompressional melting, which have been crucial for the generation of postcollisional alkaline magmas in the region. Slab roll-back has caused crustal collapse and hence extension in the upper plate, and both extension and magmatism appear to have migrated with the retreating trenches.

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