International Geology Review, Vol. 43, 2001, p. 578–594.

Introduction to the Geology of Turkey—A Synthesis

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General Statement

JUST AS ANATOLIA (Asia Minor) has had a rich cultural history as home to numerous and diverse civilizations, the geology of Turkey is a colorful, fascinating mosaic. Presently lying within the Alpine-Himalayan mountain belt near the junction of Eurasia, Africa, and Arabia, the landmass that underlies most of Turkey was once situated at the collisional boundary of two megacontinents—Gondwana in the south and Laurasia in the north. The geological framework of Turkey comprises many lithospheric fragments that were derived from the megacontinental margins and then were amalgamated during the Alpine orogeny when the Arabian plate collided with the Anatolian plate in the Late Cretaceous–Tertiary. As the rifts widened, the Tethyan oceans developed; when, subsequently, the megacontinents collided, these oceans closed sequentially.

Orogenic Belts

The age and distribution of subduction-accretion complexes and ophiolites in Turkey are consistent with the existence of two Tethyan oceans, Paleotethys and Neotethys, which partly overlap in time (e.g., Sengör, 1979a, 1987; Sengör and Yilmaz, 1981; Okay and Tüysüz, 1999; Stampfli, 2000). Although it is traditionally, for simplicity, accepted that Paleotethys was essentially a Paleozoic–Early Mesozoic ocean and that Neotethys was a Mesozoic–Early Tertiary ocean, there is no agreement on their definitions and paleogeographic locations. Three separate Neotethyan oceanic basins developed: the Intra-Pontide, northern Neotethys, and southern Neotethys (Bitlis ocean or Peri-Arabic ocean). Although the southern basin continues to survive as the eastern part of the Mediterranean Sea—an oceanic basin dating from the Triassic (e.g., Robertson, 1998) or even Late Permian (e.g., Stampfli, 2000 and references therein)—the other two no longer exist. The southern part of the country is now in a post-collisional phase (e.g., Sengör and Yilmaz, 1981; Sengör et al., 1985; Dewey et al., 1986; Pearce et al., 1990) along the Southeast Anatolian suture, whereas areas in the west, such as the Cyprus area, are still in the early collisional phase (Robertson and Grasso, 1995). These oceans closed diachronously. The northern branch of the Neotethyan ocean was closed by final collision and suturing during the Late Paleocene–late Burdigalian, whereas northward subduction across the southern branch continued through the late Middle Miocene, and then was closed entirely during the continent-continent collision across the Arabian Plate in the south and the Anatolian Plate in the north along the Southeast Anatolian suture (Yilmaz, 1993; Elmas, 1996a; Yilmaz and Yildirim, 1996). During the evolution of northern Neotethys, Turkey developed as a suite of microcontinental fragments and intervening small oceanic basins (Sengör and Yilmaz, 1981; Görür et al., 1984; Sengör et al., 1984; Robertson and Dixon, 1984). Northern Neotethys is now considered to have been a multibranched ocean, each branch of which—the Izmir-Ankara-Erzincan ocean and the Inner Tauride ocean—closed at different times (Sengör and Yilmaz, 1981; Görür et al., 1984). On the other hand, there is also evidence against the existence of the Inner Tauride ocean (e.g., Yaliniz et al. 1996 and references therein).

It is important to note that many alternative models have been proposed for the evolution of Tethyside in Turkey, and they differ in the timing of ocean-basin opening and closure, subduction polarity, and the location and number of sutures (e.g., Sengör and Yilmaz, 1981; Okay and Tüysüz, 1999; Ustaömer and Robertson, 1999; Göncüoglu et al., 2000; Stampfli, 2000; Elmas and Yigtbas, 2001 and references therein). For example, one of the main controversies concerning the Southeast Anatolian suture is its location: some suggest that the suture lies to the south of the Bitlis and Pötürge massifs (Sengör and

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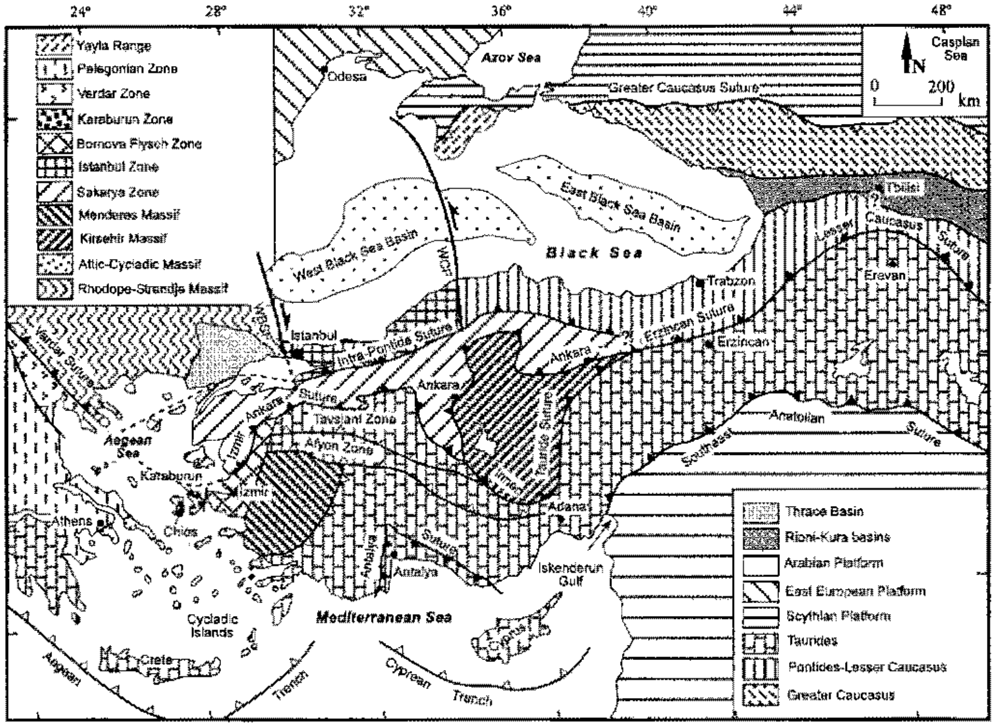

FIG. 1. Simplified tectonic map showing major neotectonic structures and neotectonic provinces (modified from Sengör et al., 1985; Barka, 1992). Abbreviations: WBSF = Western Black Sea fault; WCF = West Crimean fault. Heavy lines with filled triangles show sutures: the tips of triangles indicate polarity. Heavy lines with open triangles indicate thrust belts: triangles point toward the vergence direction. Heavy lines with half arrows are strike-slip faults: the half arrows show relative movement along these faults. The Pontides and Lesser Caucasus form the eastern extension of the Sakarya Zone.

Yilmaz, 1981; Yilmaz, 1993; Yilmaz et al., 1993) whereas others argue that the suture is located north of the massifs (Göncüoglu and Turhan, 1984; Yazgan, 1984).

In the early years, Ketin (1966) subdivided Anatolia—on the basis of geography—into the Pontides (Laurasian realm), the Anatolides, the Taurides, and the Border folds (Gondwanan realm). That proposal was followed by different subdivisions based upon the concept of plate tectonics (Wilson cycle). In particular, we choose to follow one of the most recent schemes, devised by Okay and Tüysüz (1999). In their framework, Turkey is traversed by five major Neotethyan suture zones (Izmir-Ankara-Erzincan, Intra-Pontide, Inner Tauride, Antalya, and Southeast Anatolian) that are marked by complete or partial ophiolite complexes and ophiolitic mélanges (Fig. 1).

Each of the microcontinental fragments bounded by these sutures has a complex history. In the north, the Pontides are a tectonic patchwork that have evolved since the Triassic by progressive accretion of continental terrains, with attached oceanic fragments, during the closure of the Paleotethyan and Neotethyan oceans. The Pontides also were subjected to the Cimmerian orogeny during the closure of Paleotethys and its marginal basins, the Karakaya and Küre basins (e.g., Ustaömer and Robertson, 1999; Okay, 2000 and references therein). The Pontides comprise three major tectonic zones (Okay, 1989a), namely: (1) the Strandja zone—a composite orogenic belt deformed and regionally metamorphosed during the Late Variscan (high-amphibolite facies with granitoid intrusions) and Late Jurassic–Early Cretaceous (greenschist facies) orogenies (Okay, Tansel et al., 2001); (2) the Istanbul zone with a Precambrian basement and unconformably overlying Ordovician to Eocene sediments with numerous unconformities (Aydin et al., 1986; Görür et al., 1997; Dean et al., 2000); and (3) the Sakarya zone—comprising unmetamorphosed to variably metamorphosed rocks (mainly to greenschist facies) during the Late Paleozoic and the latest Triassic, intensely deformed and imbricated basement, and Jurassic to Tertiary clastic and carbonate cover rocks (e.g., Bingöl et al., 1975; Sengör et al., 1980, 1984, 1988; Sengör and Yilmaz, 1981; Tekeli, 1981; Koçyigit, 1987, 1991a; Okay, 1989a; Altiner et al., 1991; Altiner and Koçyigit, 1993; Tüysüz and Yigtbas, 1994; Genç and Yilmaz, 1995; Yilmaz et al., 1995; Elmas, 1996b; Okay, 1996; Okay et al., 1996a; Pickett and Robertson, 1996; Göncüoglu et al., 1996–1997; Okay and Monie, 1997; Rojay and Altiner, 1998; Göncüoglu et al., 2000; Okay, 2000; Kozur et al., 2000). The most southerly of these zones, the Sakarya, is juxtaposed with the Anatolides along the main Neotethyan suture zone, the Izmir-Ankara-Erzincan zone (IAEZ).

The Anatolides and the Taurides to the south are usually treated together and, therefore, are termed the Tauride-Anatolide platform (TAP; Sengör and Yilmaz, 1981). The Anatolides comprise: (1) the Bornova flysch zone—a chaotically deformed belt consisting of large (kilometer-scale) uppermost Triassic to Cretaceous limestone blocks within a highly tectonized and intensely sheared matrix of Maastrichtian Paleocene graywacke-shale alternations (e.g., Erdogan, 1990; Okay and Siyako, 1993; Okay et al., 1996); (2) the Tavsanli zone—a belt of deformed volcano-sedimentary rocks affected by bigh pressure/low temperature metamorphism during tbe Campanian (e.g., Harris et al., 1994; Okay, 1984, 1986, 1996; Okay et al., 1998; Sherlock et al., 1999); and (3) the Afyon zone—a Paleozoic (Devonian–Permian) to Mesozoic (Triassic–Maastrichtian) sedimentary sequence consisting mainly of metaclastics and thick platform carhonates with rare metabasites, metamorphosed into greenschist facies during Paleocene time (Özcan et al., 1988; Okay, 1984; Göncüoglu et al., 1996-1997); together with (4) the Menderes Massif (MM) and the Central Anatolian Crystalline Complex (CACC).

The Taurides consist of a series of superimposed, unmetamorphosed (except for the Alanya nappes, which underwent an initial HP/LT metamorphism; Okay and Özgül, 1984; Okay, 1989b), both northerly and southerly derived nappes comprised of Lower Paleozoic to Lower Tertiary platform-type carbonates, turbidites, and continental clastic rocks (e.g., Özgül, 1976, 1985). All of these nappe piles, except for one (the Antalya Nappes), are interpreted to have been derived from the margins of Neotethyan oceanic basins to the north. Some of the nappe piles also contain Upper Cretaceous supra-subduction zone ophiolites—with their metamorphic soles—emplaced atop sedimentary thrust slices. The most internationally known pile is the Lycian Nappes (e.g., Okay, 1990; Collins and Robertson, 1999 and references therein). Thus, the TAP represents a continental platform between the Neotethyan IAE ocean to the north and the southern branch of Neotethys.

In this context, the Anatolides represent the metamorphic northern margin of the Anatolide-Tauride platform and consist of two crustal-scale, Barrovian-type metamorphic massifs, namely the Menderes Massif in the west and the CACC (Kirsehir Massif) in the east. The Menderes Massif consists of a typically thick, tripartite lithologic succession with a Precambrian gneiss basement (the so-called gneissic core), a Paleozoie "schist envelope" covering the gneiss core, and a Mesozoic–Cenozoic "marble envelope" overlying both. The massif has undergone metamorphism at greenschist- to amphibolite-facies conditions during the Middle Eocene, with the metamorphic grade increasing toward the core. The massif has a complex tectonometamorphic history involving shortening, internal imhrication, and consequent crustal thickening, erosional, and extensional exhumation, thus representing one of the world's best examples of a core complex (e.g., Bozkurt, 1996, 2001a; Bozkurt and Oherhänsli, 2001a, 2001b; Sözbilir 2001). The Central Anatolian Crystalline Complex (CACC; Göncüoglu et al., 1991) consists of medium- to high-grade metasediments (mainly marbles) of Precambrian to Late Mesozoic age and metaophiolites, with voluminous Cretaceous granitoids and associated volcanics (e.g., Akiman et al., 1993; Güleç, 1994; Yaliniz et al., 1999; Boztug, 2000; Gençalioglu-Kuscu and Floyd, 2001 and references therein). The CACC experienced regional progressive HT/MP metamorphism at greenschist-to upper amphibolite-facies conditions (with local granulite-facies conditions) in the latest Cretaceous (Göncüoglu, 1986; Seymen, 1981, 1984; Göncüoglu et al., 1991).

In southeastern Turkey, the TAP is juxtaposed with the Arabian platform (corresponding to the border folds of Ketin, 1966) along the Southeast Anatolian suture zone (e.g., Yilmaz, 1993; Yilmaz et al., 1993; Yilmaz and Yildirim, 1996; Yigitbas et al., 1996a, 1996b). The Arabian platform is made up of mostly Paleozoic–Mesozoic rocks deposited atop a craton that was stabilized during the Pan-African orogeny (e.g., Altiner, 1989). Most of the petroleum in Turkey is produced from rocks of the Arabian platform.

While the Pontides are characterized by Hercynian metamorphism-magmatism, Permo-Triassic Paleotethyan accretion-subduction complexes, and the clastic products of Liassic transgression, there are no records of such events in the Anatolides. Unlike the Pontides, the Anatolides and Taurides are characterized by relatively autochthonous Paleozoic sedimentary rocks and complex nappe structures. Alpine regional metamorphism is ubiquitous in the Anatolides but is absent in the Taurides, with only a few exceptions. During the closure of the IAE Neotethyan ocean, the TAP occupied the footwall of northward-subducting oceanic crust, whereas the Pontides comprised the hanging wall (Okay, Tansel et al., 2001 and references therein). The Anatolides were buried beneath southward-moving slices of ophiolitic and accretionary-complex material, and the northern margin was deeply subducted and enjoyed HP-LT metamorphism at depths of 50 km, thus forming the Cretaceous blueschist belt known as the Tavsanli zone. Consequently, the northern parts of the platform (Anatolides) were strongly deformed, metamorphosed, and internally sliced, whereas the Taurides comprised the cover nappes (e.g., Okay, Tansel et al., 2001 and references therein). Thus, the Pontides are devoid of nappe structure and Alpine metamorphism, and comprise a major Late Cretaceous magmatic-arc complex characterized by granitoid intrusives and widespread volcaniclastics and lava flows that are intercalated with sediments (e.g., Tüysüz et al., 1995; Çamur et al., 1996; Okay and Sahintürk, 1997; Yilmaz et al., 1997; Bektas et al., 1999; Tüysüz, 1999). The Pontide arc and related rocks are characterized by the presence of economic massive and stockwork-type polymetallic sulfide ore deposits and Kuruko-type sulfide ores.

The present configuration and mountainous nature of the Pontides and TAP have been acquired since Late Miocene–Pliocene intracontinental convergence and consequent N-S shortening. Although intracontinental convergence and N-S compression have ceased in western Turkey, these continue in eastern Turkey.

Ophiolite Belts

Isolated outcrops of Tethyan ophiolites occur throughout Turkey. Ophiolitic remnants of the Paleotethyan ocean are relatively scarce and the main exposures of such occur in the Pontides. They are commonly metamorphic and represented by a sequence of dismembered metaophiolites (serpentinized peridotite, layered cumulate gabbros, isotropic microgabbro—cut by diabase dikes—that passes into a sheeted dike complex, and basic lavas) and overlying epi-ophiolitic deep-sea cover sediments (alternations of pillow lavas, massive lava flows, lava breccias, and hemipelagic sediments, such as shales alternating with terrigenous sandstones) (e.g., the Elekdag metaophiolites, Küre complex, Karadag Nappe, Yusufeli-Artvin metaophiolites) (e.g., Koçyigit, 1991b; Yilmaz et al., 1997; Ustaömer and Robertson, 1999 and references therein). The Elekdag metaophiolite is regarded as a remnant of the Paleotethys Ocean (Sengör, 1979a, 1990; Sengör et al., 1984; Tüysüz, 1990; Ustaömer and Robertson, 1999) whereas the Küre complex is considered a subduction-accretion complex (e.g., Ustaömer and Robertson, 1997, 1999 and references therein). These ophiolitic fragments also contain evidence for blueschist- and eclogite-facies metamorphism, which probably was associated with their emplacement during the Late Triassic.

Upper Cretaceous Neotethyan ophiolites represent the variably disrupted fragments of the Neotethyan oceanic branches and record the destruction of these oceanic basins. The ophiolites occur either as dismembered allochthonous fragments/blocks within mélanges or as huge thrust sheets of unfragmented oceanic crust. Most are characterized by an epi-ophiolitic sedimentary cover dominated by pelagic and/or volcanogenic deep-sea sediments (e.g., Robertson, 1994). They occur along E-W–trending belts, corresponding to Neotethyan sutures: the northern and southern, in which the northern Neotethyan ophiolites are relatively less known compared to those of the southern Neotethys. The Izmir-Ankara-Erzincan suture in the north is represented mainly by the widespread occurrence of ophiolitic mélange. The most common and internationally known name given to this mélange is the Ankara mélange (Bailey and McCallien, 1950)—an Upper Cretaceous chaotic tectono-sedimentary ophiolitic mélange comprising various blocks of dissimilar age, origin, facies, and size set in an intensely sheared and fine-grained matrix composed mostly of ophiolitic material-rich sandstone, shale, radiolarite, and pelagic mudstone (e.g., Koçyigit, 1991b; Floyd, 1993; Bragin and Tekin, 1996, 1999). The youngest age obtained from the matrix is early Campanian (e.g., Koçyigit, 1991b; Rojay and Süzen, 1997; Tankut et al., 1998a; Bragin and Tekin, 1996, 1999). The mélange is a subduction-accretion complex that records the closure of northern Neotethys. In addition to mélanges, there are also occurrences of ophiolitic bodies, such as the Sarikaraman and Çiçekdag ophiolites in the CACC (Yaliniz et al., 1996, 1999, 2000a, 2000b; Floyd et al., 2000). Although these ophiolites occur as dismembered bodies, they retain a complete ophiolitic pseudostratigraphy. "Cyprus-type" cupriferous pyrite deposits in the Küre complex and podiform-type chromite ores in the Elekdag metaophiolite are economic aspects of the Paleotethyan ophiolites in Turkey, and significant chromite mineralization is also hosted by the Neotethyan ophiolitic rocks.

The southern belt includes well-documented ophiolite bodies. They occur either along the Southeast Anatolian suture zone (the Guleman, Gevas, Yükseova [Cilo], Berit, and Kömürhan ophiolites) or on the Taurides (the Lycian, Beysehir, Mersin, Ali Hoca, Pozanti-Karsanti, Kizildag [Hatay], Ispendere, and Antalya ophiolites). The Lycian, Beysehir, Mersin, Ali Hoca, and Pozanti-Karsanti ophiolites record ophiolite obduction derived from northern Neotethys during the Late Cretaceous, whereas the Kizildag, Ispendere, and Antalya ophiolites were derived from southern Neotethys. Where observed, the ophiolitic bodies rest tectonically on the Tauride platform carbonates and are associated with ophiolitic mélanges and metamorphic soles (Dilek and Moores, 1990; Lytwyn and Casey, 1995; Önen and Hall, 1993, 2000; Harris et al., 1994; Polat and Casey, 1995; Parlak et al., 1995; Polat et al., 1996; Dilek and Thy, 1998; Collins and Robertson, 1999; Dilek et al., 1999; Parlak and Delaloye, 1999; Parlak, 2000).

The geochemistry and structural position of the Upper Cretaceous ophiolites of Turkey suggest a suprasubduction-zone tectonic setting (e.g., Robertson, 1994; Lytwyn and Casey, 1995; Parlak et al., 1996, 2000; Yaliniz et al., 1996; Dilek et al., 1999). Most of these comprise completely preserved ophiolitic sequences, consisting of metamorphic tectonites, cumulate layered and isotropic gabbros, plagiogranite, sheeted dike complexes, pillow lavas, and an epi-ophiolitic sedimentary cover of epiclastic volcanogenic deep-sea sediments and debris flows intercalated with pelagic units.

Metamorphic Belts

Metamorphic rocks are common elements in the geology of Turkey and they are widely distributed. They form distinct belts and contain key evidence concerning the tectonic evolution of Tethyside in Turkey. The metamorphic rocks can, based on their ages and degree of main metamorphism (latest event), be classified into three major groups as Hercynian, Cimmerian, and Alpine. The former occurs only in the Pontides, whereas the latter two are characteristic features of the Anatolides of Ketin (1966). Each of these belts records remarkable differences in tectonometamorphic evolution and thus sheds light on the tectonic evolution of Anatolia.

The Hercynian metamorphics generally comprise Precambrian greenschist- to amphibolite-facies rocks and are exposed only in the western Pontides, forming the basement to Paleozoic rocks of the Istanbul zone (Bolu Massif: Ustaömer and Rogers, 1999; Yigtbas et al., 1999). Hercynian continental units of high-grade metamorphic rocks (gneiss-amphibolite alternations) in the Pontides form tectonic windows in the core of the Kazdag and Uludag massifs. Their exposures in the eastern Pontides are known as the Pulur Massif (Okay, 1996; Okay and Leven, 1996; Topuz et al., 1997). In addition, there is also evidence for Precambrian HP/LT blueschist- and eclogite-facies metamorphism in the Menderes and Bitlis-Pötürge massifs (e.g., Okay et al., 1985; Candan, 1996; Candan et al., 2001).

Cimmerian (Triassic) metamorphic rocks underlie large areas mainly in the Pontides. Their exposures can be grouped into three belts: (1) the southern belt to the north of the Izmir-Ankara-Erzincan suture comprises the Karakaya complex and some inliers of high-grade Paleozoic metamorphic rocks (the Uludag and Pulur massifs) (Okay, 1996; Okay et al., 1996; Göncüoglu et al., 2000); (2) the northern belt exposed on the Armutlu Peninsula—the Armutlu metamorphics (the Pamukova and Iznik metamorphics) (e.g., Göncüoglu and Erendil, 1990; Yilmaz, 1991–1993); and (3) the eastern belt, namely the Kargi Massif in the central Pontides, and the Tokat and Agvanis massifs in the eastern Pontides. The latter two represent the eastern extension of the Karakaya complex in the Pontides. Most Triassic massifs are characterized by widespread occurrences of greenschist-facies rocks, but there is also evidence for blueschist- and eclogite-facies metamorphism in the Elekdag and Küre ophiolites of the Kargi Massif. The Karakaya complex represents an orogeny caused by the latest Triassic, northward obduction of subduction-accretion units of Paleotethys (Tekeli, 1981; Koçyigit et al., 1991; Okay et al., 1996).

Alpine metamorphics comprise four distinct belts: (1) Cretaceous blueschist- to eclogite-facies rocks; (2) Cretaceous greenschist- to amphibolite-facies rocks; (3) Paleocene–Eocene greenschist- to amphibolite-facies rocks; and (4) Oligocene amphibolite-facies rocks. Cretaceous blueschist- to eclogite-facies metamorphics form distinct belts and are divided into three groups, namely the Tavsanli zone and Çamlica metamorphics in western Turkey, and the Alanya Massif in the western Taurides (e.g., Harris et al., 1994; Okay, 1984, 1986, 1994; Okay et al., 1998; Sherlock et al., 1999; Okay and Satir, 2000a). There is also evidence for subsequent greenschist-facies metamorphism in the latter two. Cretaceous greenschist- to amphibolite-facies rocks form two large metamorphic culminations, known as the Central Anatolian Crystalline Complex in the so-called Kirsehir Block of central Anatolia, and the Bitlis and Pötürge massifs in southeastern Anatolia. The massifs have experienced Barrovian-type dynamothermal metamorphism at greenschist- to upper amphibolite-facies conditions, and locally at granulite-facies conditions. The CACC comprises three large metamorphic bodies: the Akdagmadeni Massif in the northeast, the Nigde Massif in the south, and the intervening Kirsehir Massif (e.g., Göncüoglu et al., 1991; Whitney and Dilek, 1997, 2001; Whitney et al., 2001).

Paleocene–Eocene greenschist- to amphibolite-facies rocks occur only in western Anatolia, mainly in the Afyon zone and the Menderes Massif, respectively, and are attributed to burial at the base of the southward-moving Lycian nappes. On the other hand, Oligocene amphibolite-facies metamorphic rocks exposed in the Kazdag Massif are thought to be related to the extensional exhumation of the massif (Okay and Satir, 2000b). The Menderes Massif also experienced low-grade greenschist-facies metamorphism during the Miocene in the footwall of the presently low-angle normal faults. Both the Menderes Massif and the Kazdag Massif are interpreted as metamorphic core complexes. In particular, the Menderes Massif is the subject of much recent research because of its importance in understanding the tectonic processes involved in core-complex formation, and also because of its significance in gaining a better understanding of western Anatolian tectonics insofar as the massif forms an important bridge between the so-called paleotectonic and neotectonic periods of the region.

In addition to the above, Late Jurassic greenschist-facies rocks are the most typical element of the Strandja zone. These rocks have been named the Strandja Massif (Okay, Satir et al., 2001 and references therein).

Cenozoic Volcanism

Not surprisingly, given the tectonic history of Anatolia, there are vast areas underlain by volcanic rocks. Volcanic activity has been closely associated with the closure of oceans and with neotectonic regimes. Five main Tertiary-Quaternary volcanic provinces, namely the Western Anatolian volcanics (WAVP), Galatia volcanics (GVP), Central Anatolian volcanics (CAVP), Eastern Anatolian Volcanics (EAVP), and the Karacadag volcanics (KVP) are notable, but there are many other volcanic centers that have received the attention of researchers, and for good reason. Geologically very young, even historical, volcanism has occurred in Anatolia, such as at Nemrut Dagi in eastern Anatolia, Erciyes and Hasan Dagi in central Anatolia, and in the Kula area (where cinder locally contains supposed human footprints) of western Anatolia. Therefore, across Turkey there are volcanoes that must be classified as dormant, and accordingly there is significant potential for geothermal energy. Furthermore, unusual natural landscapes (such as fairy chimneys), ancient cave dwellings, and monuments carved out of young volcanic rocks in the Cappadocia and Afyon areas draw tourists from around the world, and the preservation of these sites falls within the purview of geoscientists (e.g., Topal and Sözmen, 2001). One of the most striking features of these volcanic provinces, particularly the CAVP and WAVP, is the presence of numerous polygenetic volcanoes and monogenetic cones scattered throughout, the distribution, location, and vent alignment of which are tectonically controlled (e.g., Toprak, 1998).

The volcanic provinces of Turkey are characterized by Neogene–Quaternary volcanic activity, mainly divided into two episodes, although the character and timing of each phase differs among the provinces. For example, the Oligocene–Early Miocene episode of extensive calc-alkaline granitoid intrusion and associated volcanism was followed by Late Miocene–Pliocene alkaline volcanism in western Anatolia (e.g., Yilmaz, 1990; Delaloye and Bingöl, 2000; Yilmaz et al., 2001 and references therein). On the other hand, volcanic activity in the EAVP is represented by Miocene–Pliocene collision-related volcanic rocks (11 ± 2.7 Ma; Keskin et al., 1998) and by Quaternary volcanism in a few localized centers, mostly following N-S tensional openings (e.g., Yilmaz et al., 1987, 1998). The last eruption in the region occurred in 1441 at Nemrut Dagi (Tchalenko, 1977).

A similar evolution can be deduced for the CAVP, a calc-alkaline province consisting mostly of an early phase of basaltic-andesitic lava flows (13.5–8.5 Ma), followed by voluminous rhyolitic ignimbrite ejection with interbedded Plinian pumice-fall deposits (9.5–9 to 2.7 Ma) (Pasquare et al., 1988; Le Pennec et al., 1994; Schumacher and Mues-Schumacher, 1996; Temel et al., 1998 and references therein). Two distinct cycles of volcanism in the GVP occurred during the Early Miocene and Late Miocene (Wilson et al., 1997; Tankut et al., 1998b), but there is also evidence that some magmatic activity in the province may be as old as Late Cretaceous (Keller et al., 1992; Koçyigit et al., 2000). The Neogene volcanic rocks locally host important gold mineralization (e.g., Bergama and Usak in western Anatolia) and, therefore, are the subjects of active exploration.

Sedimentary Basins

Throughout Turkey, the Mesozoic–Lower Tertiary deformed Neotethyan units are unconformably overlain by Neogene basins. The basins cover vast areas and the basin fill is mainly composed of Miocene fluvial-lacustrine sediments (e.g., Görür et al., 1998). In western Anatolia, Miocene sedimentation occurred in E-W–trending grabens (e.g., the Gökova, Büyük Menderes, Gediz, and Simav grabens) and approximately N-S–trending basins (e.g., the Soma, Usak-Selendi, Gördes, and Emet basins). The N-trending basins are characterized by central alkaline- and calc-alkaline volcanism. These basins possibly were formed during orogenic collapse, following post-collisional intracontinental convergence and consequent crustal thickening in western Anatolia. On the other hand, it should be emphasized here that the origin and tectonic significance of these basins and their mutual relationships are still debated. The basins of western Anatolia are economically important in that they contain copious amounts of lignite and borate minerals (e.g., Inci, 1998a, 1998b; Helvaci and Alonso, 2000 and references therein).

Miocene foreland basins occur along the Taurides and they cover unconformably the carbonate platforms and the nappe piles. Locally, the nappes are thrust onto the Miocene sediments. Some of the most important ones are the Adana, Aksu, Köprü, Manavgat, and Mut basins (e.g., Hayward, 1984; Akay et al., 1985; Robertson et al., 1991; Flecker et al., 1995, 1998; Williams et al., 1995; Yetis et al., 1995; Glover and Robertson, 1998a, 1998b; Karabiyikoglu et al., 2000; Robertson, 2000). The Miocene in Central Anatolia is represented by extensive occurrences of lacustrine-fluvial clastics, carbonates, and evaporites accumulated in the so-called Beypazari basin (e.g., Helvaci et al., 1989; Inci, 1991). The presence of economic resources, such as lignite, bituminous shale, trona, gypsum, and clay minerals, has attracted much interest to this basin. Miocene sedimentation continued in some of the basins, which initially developed as magmatic arcs (forearc and intra-arc) and collision-related basins in the Pontides (e.g., Thrace basin) (Görür and Okay, 1996) and in the Anatolides (e.g., Çankiri, Tuzgölü, Haymana, and Sivas basins) (e.g., Çiner et al., 1996; Görür et al., 1998; Kaymakçi et al., 2000; Ocakoglu, 2001). The Thrace basin hosts important natural gas occurrences (e.g., Görür and Tüysüz, 2001) while the others contain lignite, evaporites (e.g., gypsum), and clay minerals. The Tuzgölü basin also supplies a considerable amount of salt (Ocakoglu, 2001; Tekin, 2001). In addition, there are records of Miocene marine carbonate and clastic sediments in the Pontides, which are attributed to Paratethys (Görür et al., 2000). While these occurred in the west and northwest, the Bitlis ocean between Eurasia and the Arabian platform existed through the Miocene as a narrow marine basin and accommodated the sedimentation of deep-water turbidites and shallow-marine carbonates, mainly reefal limestones.

During the late Middle Miocene, the total elimination of the Bitlis ocean and consequent terminal suturing of Arabia with Anatolia along the Southeast Anatolian suture led to long-lasting continental collision and intracontinental convergence that resulted in crustal thickening and uplift of the eastern Anatolian plateau. This timing coincided with the extensive volcanism that produced the Eastern Anatolian Volcanic Province. Moreover, Upper Miocene to Quaternary fluvial sedimentation

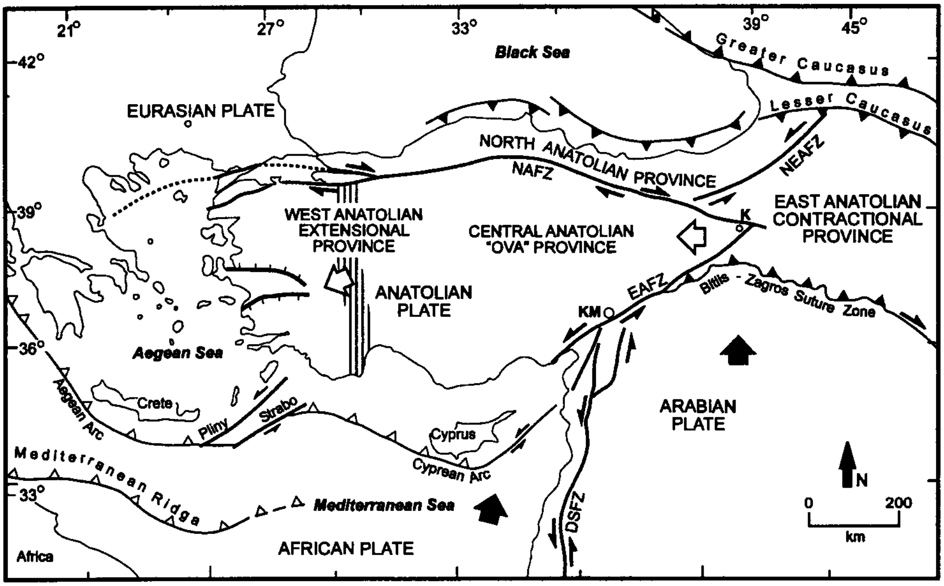

FIG. 2. Simplified tectonic map of Turkey showing major neotectonic structures and neotectonic provinces (from Bozkurt, 2001b). Abbreviations: K = Karliova; KM = Kahramanmaras; DSFZ = Dead Sea fault zone; EAFZ = East Anatolian fault zone, NAFZ = North Anatolian fault zone; NEAFZ = Northeast Anatolian fault zone. Heavy lines with half arrows are strike-slip faults. Half arrows show relative movement sense. Heavy lines with filled triangles show major fold and thrust belt; tips of the small triangles indicate direction of vergence. Heavy lines with open triangles indicate an active subduction zone; tips of the small triangles indicate polarity. The heavy hachured lines show normal faults; hachure indicates downthrown side. Bold filled arrows indicate relative movement direction of African and Arabian plates; open arrows indicate relative motion of Anatolian Plate. The hatched area shows the transition zone between the western Anatolian extensional province and the central Anatolian "ova" province (from Sengör et al., 1985).

occurred in a number of ramp basins (e.g., Mus basin). Following the end of intracontinental collision across the suture zone, the earlier compressional-contractional tectonic regime of Eastern Anatolia was replaced by a new compressional-extensional tectonic regime—the neotectonic period (with tectonic escape/extrusion tectonics)—by the early Early Pliocene.

Neotectonic Belts

Turkey is located on the seismically active Mediterranean Earthquake Belt. The country is one of the most actively deforming regions in the world and has a long history of devastating earthquakes, as reminded by the recent August 17, 1999 Kocaeli (M = 7.4) and November 12, 1999 Düzce (M = 7.2) events (e.g., Barka, 1999). The active tectonics of Turkey result from the continental collision of the African and Eurasian plates, as expressed by collisional intracontinental convergence and tectonic-escape-related deformation since the Early Pliocene (~5 Ma).

The neotectonic framework of Turkey is outlined by three major structures: the two intracontinental transform faults, namely the dextral North Anatolian Fault Zone (NAFZ; Ketin, 1969) and the sinistral East Anatolian Fault Zone (EAFZ), and the Aegean-Cyprian Arc—a convergent plate boundary where the African plate to the south is subducting beneath the Anatolian plate to the north. Furthermore, the sinistral Dead Sea fault zone also plays an important role (Fig. 2). The two strike-slip faults meet and form a continental triple junction in northeastern Turkey (e.g., Karig and Kozlu, 1990). The intervening Anatolian wedge of amalgamated crustal fragments is moving westward onto the eastern Mediterranean lithosphere, and this westward extrusion of Anatolia is accompanied by counterclockwise rotation (Rotstein, 1984). The continuum of deformation along the NAFZ and EAFZ and the westward extrusion of Anatolia have been accommodated through the internal deformation of Anatolia. Consequently, four distinct neotectonic provinces, each of which is characterized by unique structural elements and associated basin formation, have been generated: the East Anatolian contractional, the North Anatolian, the Central Anatolian "Ova," and the West Anatolian extensional provinces (Sengör et al., 1985; Fig. 2).

The pre-existing fabrics of the Tethyan evolution during the Late Paleozoic–Miocene have played important roles in directing the neotectonic structures. Of particular importance is the Western Anatolian extensional province—a region of approximately E-W–trending grabens and associated horsts. The origin and age of extension in the region has been the subject of long-lasting debate. Similarly, age and amount of total offset, and the cause of motion along the NAFZ and the EAFZ are controversial and various models have been proposed (e.g., Sengör, 1979b; Sengör et al., 1985; Koçyigit, 1989; Barka et al., 2000). Interested readers are referred to Bozkurt (2001b, 2001c) for a review. Many pull-apart basins developed along the strike-slip faults and all accommodated continental sedimentation (e.g., Koçyigit, 1987, 1989; Bozkurt and Koçyigit, 1996; Barka et al., 2000). As it is commonly believed that the NAFZ followed the Izmir-Ankara-Erzincan suture, most of the pull-apart basins were superimposed on molasse basins (e.g., Koçyigit, 1996). Conjugate strike-slip faults of dextral and sinistral character that parallel the NAFZ and EAFZ, and N-S–trending fissures, are the common structural elements of the East Anatolian contractional province. These fissures localized volcanoes (e.g., Nemrut, Süphan, and Agri) and are responsible for the widespread Plio-Quaternary volcanic rocks of the EAVP. The Central Anatolian wedge (Fig. 2) is a region where discrete pieces of continental lithosphere deformed internally along new structures or reactivated old structures. The region is characterized by mostly dextral and sinistral strike-slip faults (e.g., Ezinepazari, Almus, Tuzgölü, and Malatya-Ovacik fault zones) and extensional basins, called "ovas" (Sengör et al., 1985) that are bounded by oblique-slip faults.

Summary

In conclusion, Anatolia forms a superb laboratory for the study of subduction, ophiolite obduction, continent-continent collision, metamorphism, the relationship between lithospheric deformation and magmatism, post-collisional intracontinental convergence- and tectonic escape-related deformation and magmatism, and the consequent structures that include fold and thrust belts, suture zones, active strike-slip faulting, and active normal faulting and associated basin formation. Those interested in other recent treatments of Turkish geology may refer to the following sources: Tekeli and Göncüoglu (1984), Sengör (1989), Kelling et al. (1993), Sengör and Tatar (1996), Robinson (1997), Gourgaud (1998), Robertson and Comas (1998), Tekeli (1998), Bozkurt and Rowbotham (1999a, 1999b), and Bozkurt et al. (2000).

Acknowledgments

We thank Bora F. Rojay for his comments and fruitful discussions that improved the text.

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