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Structure and tectonic evolution of the Anatolian plateau in eastern Turkey

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ABSTRACT

The Cenozoic geology and the present lithospheric and upper-mantle structure of the Anatolian plateau in eastern Turkey and nearby regions are the result of the final collision and suturing of the continental Arabia plate to the Turkish terranes (i.e., microcontinents). This process of collision and suturing was strongly influenced by three active structures in the region: the Caucasus mountains, the Aegean subduction zone, and the Dead Sea fault system. Understanding these three major tectonic elements is important for the development of a robust model for the formation of the Anatolian plateau.

We show that the Anatolian plateau lithosphere in eastern Turkey has no lithospheric mantle, i.e., the crust floats on a partially molten asthenosphere. The average thickness of the crust in the region is ~45 km. The uppermost mantle beneath this crustal block strongly attenuates Sn waves and has one of the lowest Pn velocities on earth (~7.6 km/s). The Anatolian plateau, with an average of 2-km elevation, is dissected by numerous active seismogenic faults (mostly strike-slip and some thrust-type). Neogene and Quaternary volcanism with varying composition is widespread and covers more than half of the region.

We argue that the northward subduction of the northern and the southern branches of the Neo-Tethyan oceanic lithosphere since the Mesozoic has resulted in the development of arc and back-arc volcanism (i.e., the Pontide and Bitlis systems) and the development of the eastern Anatolian accretionary complex, which covers a large area of eastern Turkey. The northward subduction of the southern Neo-Tethys considerably thinned and weakened the overriding Eurasia plate above the descending oceanic lithosphere of the Arabia plate. The final suturing of the continental Arabia plate with the Turkish terranes in the Miocene and the continued convergence of Arabia relative to Eurasia has resulted in the shortening of the accretionary complex in both the forearc and the back-arc regions and the development of a broad zone with numerous strike-slip faults. The mobilization of the Caucasus is also partially a con-

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sequence of this convergence. The documented major episode of widespread volcanism at ca. 11 Ma is probably related to the break-off of the shallowly descending oceanic segment of the Arabian lithosphere beneath eastern Turkey. The continued convergence of Arabia relative to Eurasia resulted in the development of the North Anatolian fault and subsequently the East Anatolian fault in the Pliocene. At about this time, the northern segment of the Dead Sea fault also developed in Lebanon and northwest Syria and joined the East Anatolian fault to form the Anatolian-Arabian-African triple junction in the Maras region of southern Turkey. The development of these fault systems (i.e., North Anatolian fault, East Anatolian fault, and Dead Sea fault) provided the mechanism for the tectonic escape of the Anatolian crustal block toward the Aegean arc system.

Keywords: Anatolian plateau, eastern Turkey, lithosphere structure, Arabia plate

INTRODUCTION

This article is a brief summary of the most current models for the geologic and tectonic evolution of the Anatolian plateau in eastern Turkey and the surrounding regions (Fig. 1). Such a summary is warranted in light of the seismological and other geophysical information that became available during the past few years concerning the upper mantle and lithospheric images (e.g., Al-Lazki et al., 2003; Gok et al., 2003; Turkelli et al., 2003; Zor et al., 2003; Maggi and Priestley, 2005). This is not a review-type article; its main focus is on further advancing the understanding of eastern Turkey using and building on the few available synthesis papers, mostly based on surface geology. Recently, Şengör et al. (2003) and Keskin (2003, 2005) published excellent syntheses of eastern Anatolia and proposed similar evolutionary models. Their models are mostly based on surface geology and magma genesis, but are also consistent with recent seismological observations. In this article, we present some modifications on the proposed models of Şengör et al. and Keskin. In addition, we call attention to the importance of regional tectonic structures in the geologic evolution of the Anatolian plateau.

WHAT DO WE KNOW ABOUT THE EASTERN ANATOLIAN PLATEAU?

Crustal and Upper Mantle Structure

Excellent summaries of the geology, tectonic history, and the magma genesis of eastern Anatolia are published in the literature (e.g., Şengör and Yılmaz, 1981; Dewey et al., 1986; Dilek and Moores, 1990; Pearce et al., 1990; Yılmaz, 1993; Dilek et al., 1999; Bozkurt, 2001; Gorur and Tuysuz, 2001; Şengör et al., 2003; Keskin, 2003, 2005). It is not necessary to repeat such information in this short article. However, we provide a brief summary of the recent results based on the Eastern Turkey Seismic Experiment (ETSE). During the experiment we deployed a twenty-nine PASSCAL broadband seismic network from October 1999 to August 2001 in eastern Turkey in order to obtain detailed images of the lithospheric and upper-mantle seismic velocity structure (Sandvol et al., 2003a). These were the first detailed images of the eastern Turkey subsurface.

One of the important results was that the crustal thickness is, on average, less than 45 km (Zor et al., 2003). The crustal thickness varies from ~38 km in the Arabian foreland in southern Turkey to ~50 km farther north in the mountain ranges that extend along the Black Sea (i.e., the Pontides). The region is seismically very active, with numerous microearthquakes occurring daily. Though the majority of well-located events correlate well with mapped seismogenic faults (Fig. 2), there are many events that occur in areas where no surface faults are mapped (Turkelli et al., 2003). Most events occur in the upper 25 km of the crust, and no subcrustal earthquakes occur anywhere in the region. Most focal mechanisms are of the strike-slip type (Fig. 2), with very limited thrust or normal fault types that are restricted to certain areas (Orgulu et al., 2003).

The uppermost mantle beneath eastern Turkey is extremely anomalous. Sn waves are not observed across the Anatolian plateau (Fig. 3) and are shown to be blocked over very short distances (Gok et al., 2003; Al-Damegh et al., 2004). The tomographic Pn velocity indicates that the Anatolian lithospheric mantle is seismically very slow (7.6–7.8 km/s; Fig. 3) (Al-Lazki et al., 2003, 2004). Recent results based on surface waveform tomography also show a low shear wave velocity anomaly in the uppermost mantle beneath the Anatolian plateau (Maggi and Priestley, 2005). These independent measurements indicate both that the uppermost mantle is partially molten (i.e., there is no mantle lid) and that the asthenospheric material is in direct contact with the base of the crust. These results also indicate that continental Arabia is not significantly underthrusting beneath the Anatolian plateau, in contrast to the situation in the Indian-Eurasian collisional belt. Recent results (Sandvol and Zor, 2004) indicate that the ultra-low Pn velocity zone is also underlain by a slightly low-velocity zone in the upper mantle beneath the northern Arabia plate and the easternmost portion of the Anatolia plate. Finally, shear wave splitting results (Fig. 4B)

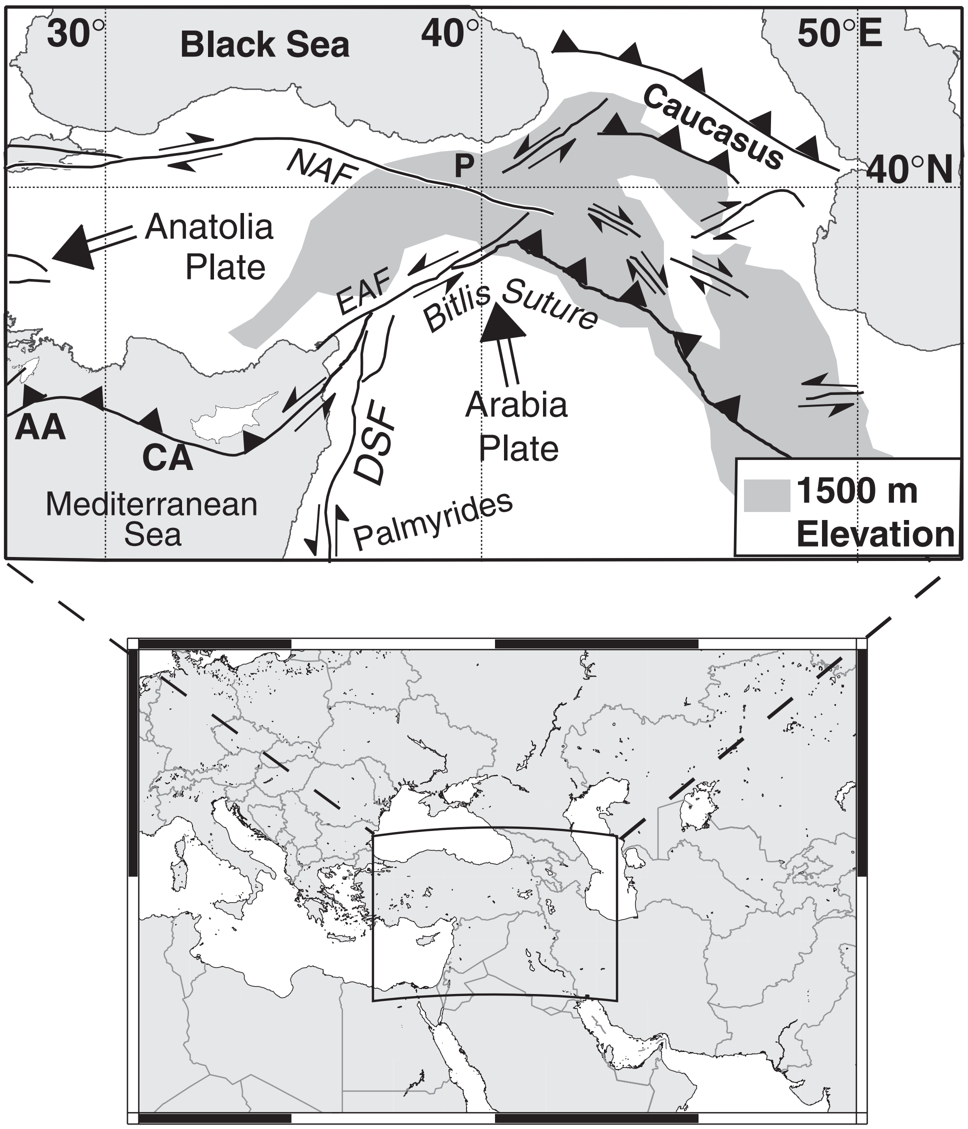

Figure 1. Map showing the major tectonic boundaries in eastern Turkey and the surrounding regions. NAF—North Anatolian fault; P—Pontide belt; EAF—East Anatolian fault; AA—Aegean arc; CA—Cyprian arc; DSF—Dead Sea fault. The large arrows indicate the direction of plate motion.

show that the fast polarization directions are uniform and exhibit northeast-southwest orientations (Sandvol et al., 2003b). These observations are best interpreted to indicate a northeast-oriented asthenospheric flow in the upper mantle that is nearly parallel to the inferred absolute plate motion.

3-D Gravity Modeling of Eastern Anatolia

To further test this hypothesis that the crust in eastern Turkey is underlain by hot, low-density material, we have studied the 3-D gravity anomalies in the region (Seber et al., 2001). If the hot (i.e., low-density) material is at sub-Moho depths, observed gravity anomalies can be used to infer the presence and perhaps the thickness of this asthenospheric material. For this purpose we calculated a new gravity map published by Ates et al. (1999). Their data show that Bouguer gravity anomalies in eastern Turkey are as low as –160 mgal. As described earlier, a recent study by Zor et al. (2003) mapped the Moho thickness in the region using seismic receiver functions from the twenty-nine stations of the ETSE seismic experiment (Fig. 5). We used their depth-to-Moho values, then gridded and extrapolated them to obtain a uniform Moho depth coverage (Fig. 5). We then calculated the 3-D gravity anomaly using this Moho structure. In our calculations we used an average of 0.5 g/cm^3 density contrast between the Moho and the sub-Moho material. This is an average density contrast and has been used in many other studies (e.g., Seber et al., 2001). We subtracted the observed gravity values from the calculated ones to obtain a residual gravity map of the region. This residual map (Fig. 6) shows that in the western part of the plateau a reasonable match between observed and calculated values is obtained. The residuals are on the order of 10–15 mgal. However, in the eastern section the amplitudes

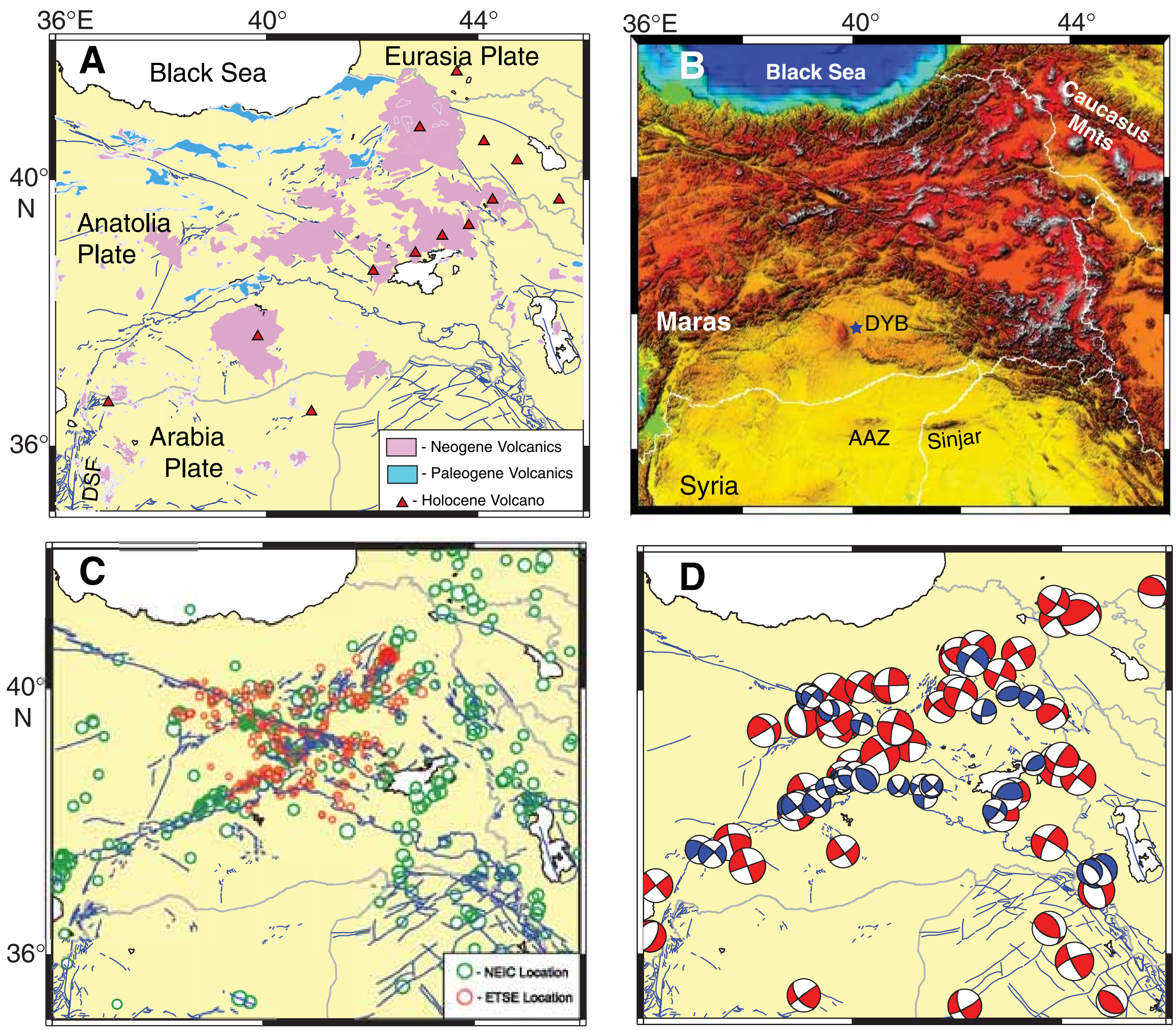

Figure 2. (A) Map showing faults and the Cenozoic volcanic rocks and Holocene volcanoes in eastern Turkey and the surrounding regions. (B) Topographic map of eastern Turkey and surrounding regions. DYB—Diyarbakir; AAZ—Abdul-Aziz uplift. (C) Seismicity map of eastern Turkey. The green circles indicate earthquake epicenters taken from the National Earthquake Information Center (NEIC) for the period 1990–2004; the red circles indicate all of the 470 well-located epicenters from the Eastern Turkey Seismic Experiment (ETSE) (Turkelli et al., 2003). The ETSE was conducted from October 1999 to August 2001. (D) Map showing the focal mechanisms for events in the Anatolian plateau and nearby regions. The red focal spheres are taken from the Harvard Centroid Moment Tensor catalog. The blue focal spheres indicate fault plane solutions obtained from the moment tensor inversion and first motion analysis using data recorded during the ETSE (Orgulu et al., 2003).

of the residuals are as high as 80 mgal. Undoubtedly, local basins and near-surface materials impact the observed gravity values; however, their amplitude and anomaly length do not have a significant impact on this study, because we are interested only in large-scale, bulk variations in gravity anomalies. Gravity anomalies reaching amplitudes as low as –80 mgal in the east need to be explained.

Two potential sources for this residual anomaly exist. If the crustal density values in the region change laterally from west to east, it is plausible that crustal density variations are the main source of these low gravity anomalies. However, because we have results from other geophysical studies, such as very slow Pn velocities and very high Sn attenuation in the uppermost mantle (Fig. 3) and no indication of large variations in bulk crustal properties from east to west, we interpret these low gravity residuals as resulting from the presence of asthenopsheric material at sub-Moho depths. With the assumption that the residual anomalies result from rising asthenospheric material at shallower depths, we can estimate the potential thickness of this layer. Our study shows that if we make an assumption that the asthenospheric material has a difference of density of –0.05 g/cm^3 compared to its surroundings, we estimate the base of this anomaly at ~70 km of depth. However, because the area of our study is relatively small, there is a possibility that as we go deeper our model loses its resolution.

Like other geophysical studies published earlier, this gravity study also supports the presence of asthenospheric material underlying the Moho in this region. These results are meant to

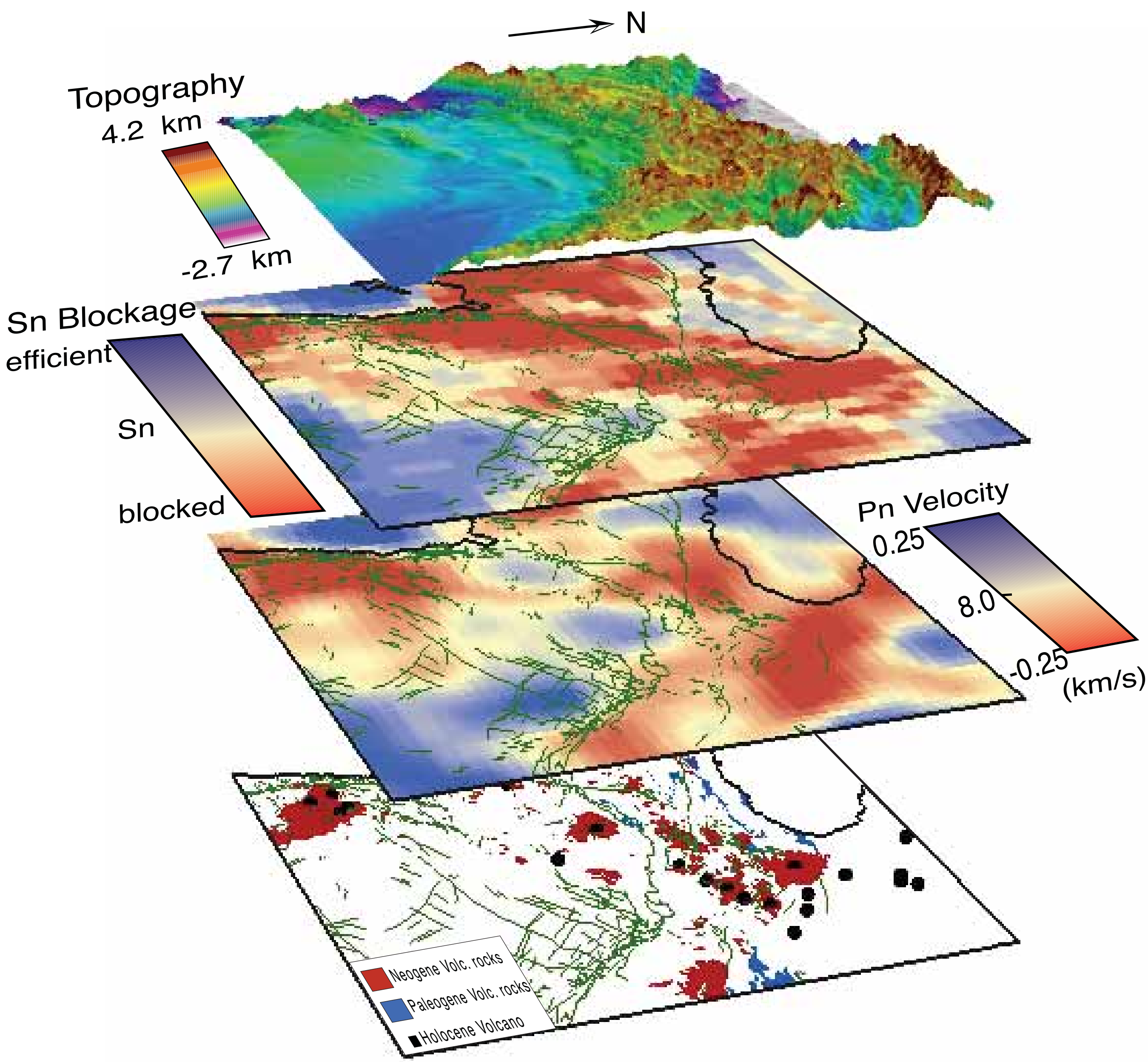

Figure 3. A perspective view with four different layers starting with a topographic map as the top layer. The second layer from the top is a map of Sn blockage with the same geographic coverage as shown in the topographic map (Al-Damegh et al., 2004). The red regions are regions of Sn blockage, and the blue regions are regions of efficient Sn propagation. The third layer from the top is a map of Pn velocities; the red regions are where the lithospheric mantle is seismically slow (~7.7 km/s), and the blue regions are where the lithospheric mantle has normal seismic velocities (~8.1 km/s) (Al-Lazki et al., 2004). The bottom layer shows the Cenozoic volcanic rocks in the Anatolian plateau and surrounding regions as well as the Holocene volcanoes (see also Fig. 2A).

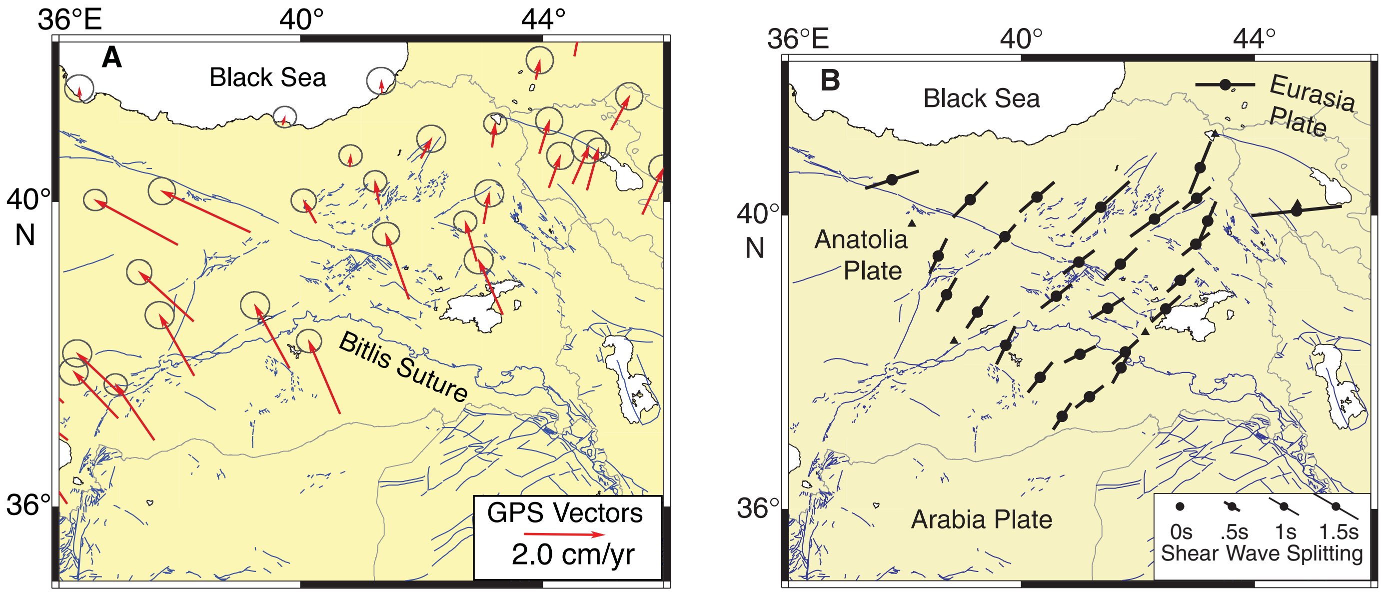

Figure 4. (A) Map showing GPS velocity vectors in the Anatolian plateau and surrounding regions from McClusky et al. (2000). (B) Map showing averaged fast directions and delay times from SKS and SKKS shear wave splitting measurements taken from Eastern Turkey Seismic Experiment broadband waveforms (Sandvol et al., 2003b). Fast directions were found to be primarily northeast.

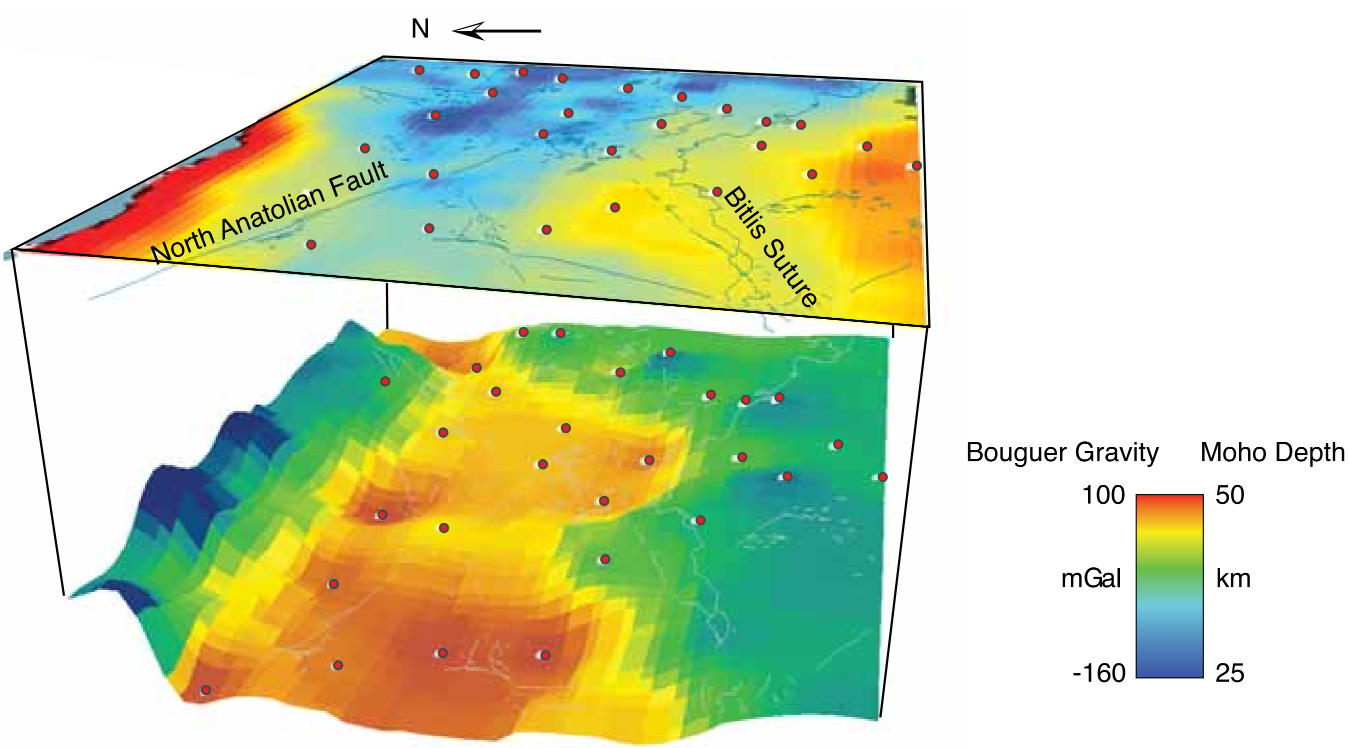

Figure 5. A perspective view of the 3-D Moho model and the Bouguer gravity anomaly map of eastern Anatolia and nearby regions. The top map image shows variations in Bouguer gravity anomalies (Ates et al., 1999). The image below shows the variations in Moho depth in the region as obtained by Zor et al. (2003). For reference, the main faults and seismic stations of the Eastern Turkey Seismic Experiment are overlain on both images (red circles).

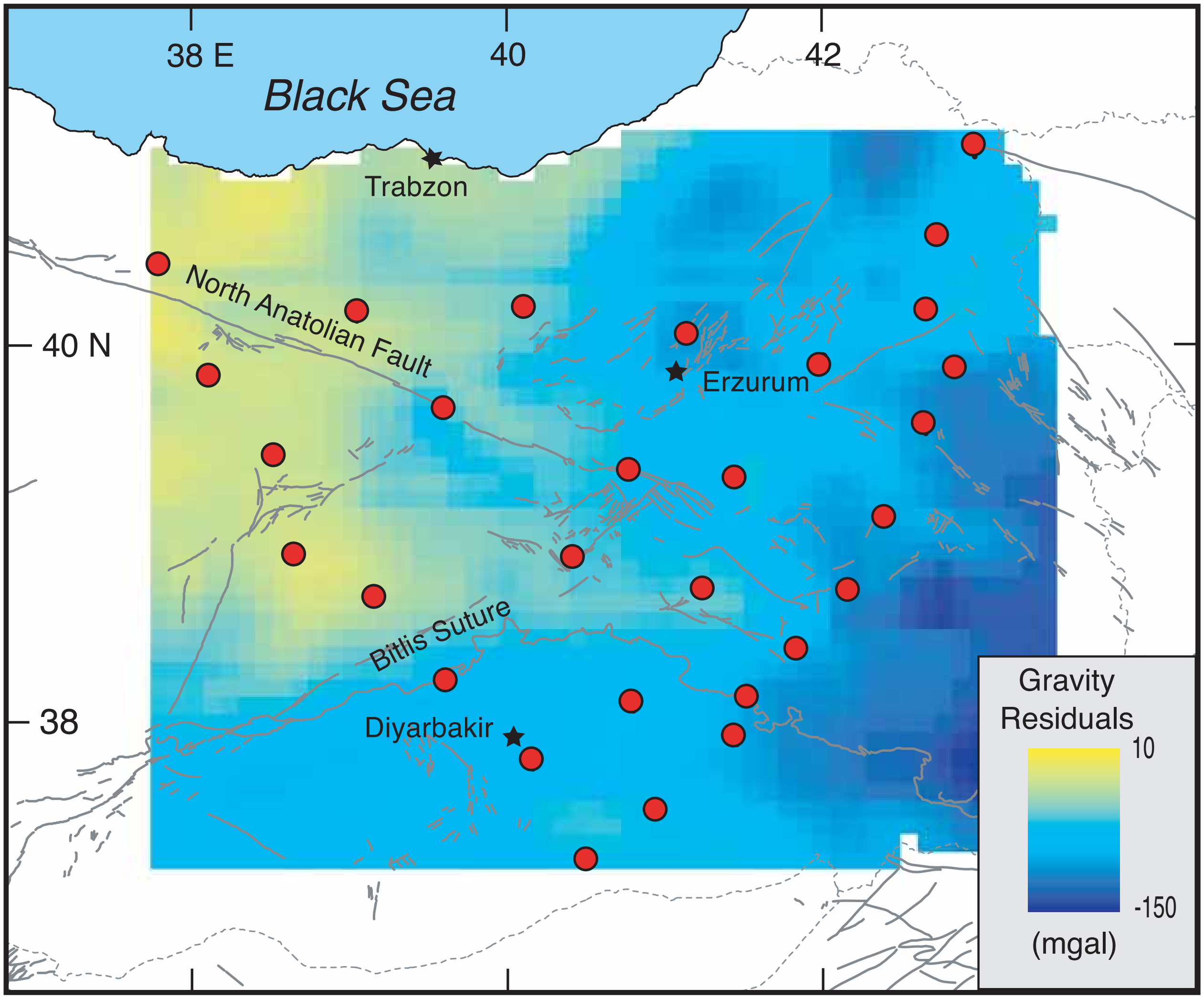

Figure 6. Map showing 3-D gravity residuals calculated using the Moho depth image shown in Figure 5. The red circles indicate seismic station locations for which Moho depth estimates are available (Zor et al., 2003). Low gravity residuals in the eastern Anatolian plateau and south of the Bitlis suture are interpreted to indicate lower-density material at sub-Moho depths.

provide first-order estimations; future work studying the region's gravity anomalies in the light of new seismological data is inevitable and could provide additional constraints in the region.

In summary, the Anatolian plateau in eastern Turkey is an elevated crustal block that is floating on partially molten hot asthenospheric material. The 2-km average elevation of the plateau is not supported by simple Airy isostatic compensation; rather, the plateau is supported by low-density asthenospheric material (e.g., Şengör et al., 2003). Furthermore, the crustal block is dissected by a mosaic of mostly seismogenic strike-slip faults, but few active compressional structures. As important, about half of the area of the plateau is covered with late Cenozoic volcanic rocks of diverse composition (Dilek and Moores, 1990; Pearce et al., 1990; Keskin, 2003). Clearly, the crust of eastern Turkey is hot and weak and is composed of crustal slivers that are in motion relative to one another. In some respect this is similar to the crustal mosaic of central Iran (e.g., Şengör and Kidd, 1979). Unlike in Iran, however, there is no well-developed fold and thrust belt in front of the Bitlis suture similar to that in the Zagros. This may reflect lateral variations in the nature of the passive continental margin of the leading edge of the colliding Arabia continental plate and/or variations in the Iranian versus the Turkish terranes and their ability to partially escape from advancing Arabia.

It is important to emphasize here that the overall structure of the Anatolian plateau, as discussed earlier, precludes the occurrence of very large earthquakes in the region, and that the maximum possible earthquake magnitude is probably ~7.

A NEW MODEL FOR THE CENOZOIC EVOLUTION OF EASTERN ANATOLIA

The complex geology of the Anatolian region of eastern Turkey is a result of the region's being located between two convergent supercontinents, Gondwana in the south and Eurasia in the north. Throughout the Mesozoic and most of the Cenozoic, these two landmasses were separated by the Paleo-Tethys and, subsequently, the Neo-Tethys (e.g., Şengör and Yılmaz, 1981; Dilek and Moores, 1990; Stampfli et al., 2001). Continental fragments of varied geologic history were also present between the two supercontinents and were accreted to Eurasia throughout Mesozoic and Cenozoic time. The northern oceanic branch of the Neo-Tethys was subducting beneath the southern margin of Eurasia with the associated arc and back-arc volcanism (Fig. 7) and was finally closed by collision and suturing during the late Paleocene along the Pontide arc (e.g., Şengör and Yılmaz, 1981; Dewey et al., 1986; Bozkurt and Mittwede, 2001). However, a segment of the southern oceanic branch of the Neo-Tethys (i.e., the Arabia plate) continued its northward subduction beneath eastern Anatolia through the middle Miocene (Fig. 7) (e.g., Yılmaz, 1993). The Bitlis arc is an integral part of this subduction system. We infer that this subduction episode throughout the Paleogene and the first half of the Neogene considerably weakened and thinned the upper plate by thermal processes and was probably similar to what happened during the early Cenozoic in western North America (e.g., Scholz et al., 1971; Hyndman et al., 2005) or beneath many back-arc basins located behind the island arcs of the western Pacific (e.g., Karig, 1971; Barazangi et al., 1975; Taylor, 1995). This upper plate is mostly the site of the East Anatolian accretionary complex, which was associated with the Pontide subduction system (e.g., Şengör et al., 2003). An important observation is the near-absence of any volcanism along the Bitlis belt and the back-arc region during most of the Paleogene and the first half of the Neogene (until ca. 11 Ma) (Fig. 7). This observation suggests that the descending oceanic part of the Arabia plate had a very shallow angle of subduction, possibly similar to what is observed in the Sierras Pampeanas in the Andes today (e.g., Barazangi and Isacks, 1976). The continued convergence of the Afro-Arabia plate relative to Eurasia produced considerable deformation of the accretionary complex in the forearc and the back-arc regions (e.g., Yılmaz, 1993). It appears that the compressive stress field that was generated by this convergence was transmitted farther inland in the upper plate and reactivated the deformation and uplift of the Caucasus intracontinental mountain belts (e.g., Philip et al., 1989; Barka and Reilinger, 1997; Reilinger et al., 1997; Saintot and Angelier, 2002).

The final continental collision and suturing of Arabia with the Turkish microplate in the middle Miocene resulted in the complete annihilation of subduction of the Neo-Tethyan ocean in eastern Turkey (e.g., Şengör and Kidd, 1979; Dewey et al., 1986; Yılmaz, 1993; Bozkurt and Mittwede, 2001; Şengör et al., 2003). Subsequent to this collision, a major episode of widespread volcanism with varied and complex composition affected most of the eastern Anatolian region starting at ca. 11 Ma (e.g., Keskin, 2003, 2005). This is a significant observation, for the present structure of eastern Anatolia has, to a large extent, been shaped by this episode. Recently, Şengör et al. (2003) and Keskin (2003) published a synthesis of the geologic evolution of and magma genesis within eastern Anatolia. Their proposed models are very similar. Şengör et al. (2003) proposed that break-off of a northward subducted slab was the key tectonic event that allowed hot, partially molten asthenospheric material to be in direct contact with the bottom of the crust. This could explain the extensive melting, the initiation of collisional volcanism, and the relatively rapid regional uplift to form the 2-km-high eastern Anatolian plateau. Our seismological results based on the recent ETSE experiment in eastern Turkey strongly support this proposed scenario (Gok et al., 2003; Al-Lazki et al., 2003, 2004; Al-Damegh et al., 2004). Both, Şengör et al. and Keskin proposed that the detached slab is the one that descended beneath the Pontide arc system, i.e., the northern oceanic branch of the Neo-Tethys.

We propose an alternative scenario: that the detached slab is the southern oceanic Neo-Tethys that was shallowly descending beneath the Bitlis arc system, i.e., the oceanic segment of the Arabian lithosphere (Fig. 7). This slab break-off and the subsequent sinking of the detached slab into the upper mantle was the main cause of the widespread Anatolian episode of volcanism and the dynamically supported uplift of the eastern Anatolian plateau. Both the initiation of the volcanism episode and the uplift have been dated at ca. 11 Ma (e.g., Pearce et al., 1990; Şengör et al., 2003; Keskin, 2003, 2005).

The convergence of Arabia relative to Eurasia continued throughout Miocene and Quaternary time and is still very active throughout eastern Turkey as evidenced by the occurrence of earthquakes, historical volcanoes, and GPS measurements (e.g., McClusky et al., 2000; Fig. 4A). This convergence has resulted in the development of compressional structures, including many east-west-striking folds and thrust faults and numerous dextral and sinistral conjugate strike-slip faults (e.g., Şengör et al., 1985; Hempton, 1987; Barka and Kadinsky-Cade, 1988; Dilek and Moores, 1990; Bozkurt, 2001). Numerous, though relatively minor, pull-apart basins have also developed that are clearly related to movement along the associated strike-slip faults (Bozkurt, 2001). Furthermore, this ongoing oblique convergence of Arabia is being accommodated partially by movement along the conjugate strike-slip faults in eastern Anatolia and partially by pure shortening in the Caucasus by means of thrusting, i.e., strain partitioning (Jackson, 1992).

The next and the final major tectonic events in the late Cenozoic geologic evolution of Anatolia were the development of the North Anatolian fault, the East Anatolian fault, and the Dead Sea fault. The Dead Sea fault is an integral part of the East Africa–Red Sea rift system. The development of a mature seafloor-spreading episode along the Red Sea in the early Miocene signaled the separation of Arabia from Africa and the development of the Dead Sea transform plate boundary farther to the north (e.g., Garfunkel, 1981; Hempton, 1987; Martinez

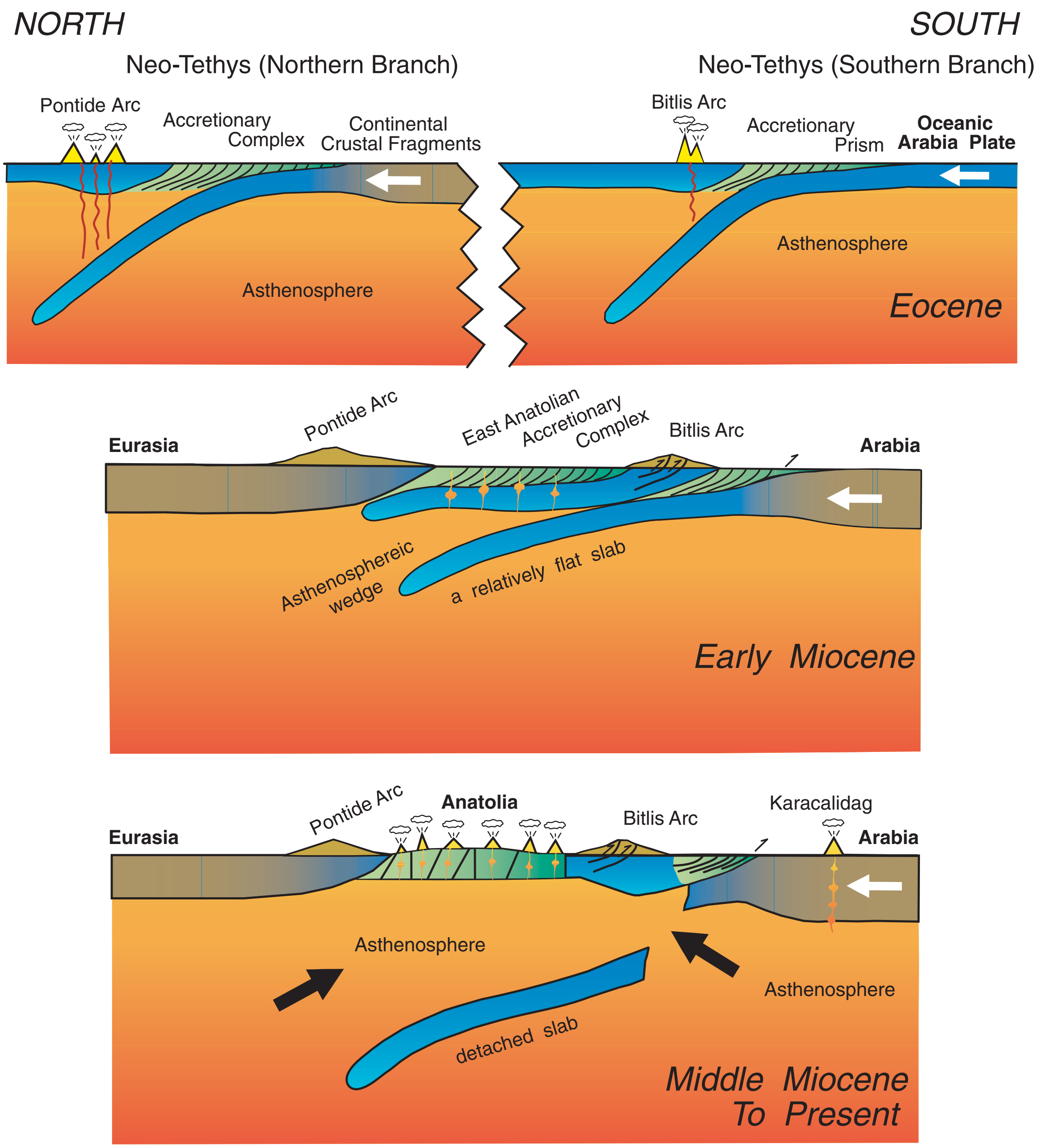

Figure 7. Schematic cross-sections showing the tectonic evolution of the Anatolian plateau in eastern Anatolia from the Eocene to the present. The cross-sections are constructed approximately along the 42°E longitude line.

and Cochran, 1988). Geologic evidence, however, indicates that the northern branch of the Dead Sea fault in Lebanon and northwest Syria was developed in only early Pliocene time (e.g., Chaimov et al., 1990; Brew et al., 2001). The resulting northward differential relative plate motion between Arabia and Africa accelerated the convergence of Arabia relative to Eurasia in the early Pliocene (e.g., Bozkurt, 2001). This apparently led to the development of the North Anatolian fault in the early Pliocene and, subsequently, the East Anatolian fault (e.g., Bozkurt, 2001). The northernmost segment of the Dead Sea fault joined the East Anatolian fault in the Maras region of southern Turkey to form the Anatolian-Arabian-African triple junction (e.g., Karig and Kozlu, 1990; Perincek and Cemen, 1990). The development of these fault systems—i.e., the North Anatolian fault, East Anatolian fault, and Dead Sea fault—provided the mechanism for the tectonic escape of the Anatolian crustal block toward the Aegean arc system (e.g., Burke and Şengör, 1986). It is of interest to note here that both the North

Anatolian fault and the Dead Sea fault appear to have formed after the thermal weakening of the Anatolia and the Arabia plates, respectively. Both fault systems, moreover, formed and propagated near the boundary of a stable and probably strong oceanic lithosphere (i.e., that of the Black Sea and the eastern Mediterranean for the North Anatolian fault and the Dead Sea fault, respectively) and a weak, thermally eroded continental lithosphere.

The story of the geologic evolution of Anatolia, as discussed earlier, also significantly affected the late Cenozoic evolution and active tectonics of the northern Arabian platform to the south of the Bitlis suture, especially in Syria (e.g., Barazangi et al., 1993). The compressive stress regime that is associated with the Arabia-Anatolia continental collision has apparently been partially transferred to the interior of the continental Arabia plate and has resulted in the crustal shortening of a few weak intraplate geologic structures. In particular, the Plio-Quaternary uplifts of the Abdul Aziz and Sinjar in northern Syria and Iraq, respectively (e.g., Brew et al., 1999), and the accelerated deformation and uplift of the Palmyra fold and thrust belt in central Syria (e.g., Sawaf et al., 2001) are best explained as a result of the collision process.

CONCLUSIONS

We show that the eastern Anatolian plateau is underlain by a crust with an average thickness of 45 km that is directly floating on a partially molten asthenosphere (i.e., there is no lithospheric mantle). The plateau, with an average 2-km elevation, is dissected by numerous active seismogenic faults with widespread Neogene and Quaternary volcanic rocks.

We argue that the northward subduction of the southern branch of the Neo-Tethys oceanic lithosphere (i.e., the Arabia plate) has considerably thinned and weakened the overriding Eurasia plate (i.e., the Anatolia microplate). The final suturing of the continental Arabia plate with the Anatolia microplate in the Miocene and the continued convergence of Arabia relative to Eurasia has resulted in the intense deformation of Anatolia and in the partial mobilization of the Caucasus. Because of the relative lack of volcanism during late Eocene to Miocene time, we suggest that the descending oceanic segment of the Arabian lithosphere was at a very shallow angle. The well-documented major episode of widespread volcanism at ca. 11 Ma is probably related to the break-off of this shallow descending plate. This could explain the extensive melting, the initiation of collisional volcanism, and the relatively rapid regional uplift to form the 2-km-high eastern Anatolian plateau.

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