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EAST ANATOLIAN HIGH PLATEAU AS A MANTLE-SUPPORTED, NORTH-SOUTH SHORTENED DOMAL STRUCTURE

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ABSTRACT

The East Anatolian High Plateau is a region of average ~2 km elevation a.s.l. exhibiting active diffuse N-S shortening and widespread Pliocene to recent volcanicity. Its elevation was hitherto thought to result from a presumed crustal thickness of ±55 km. Seismic data collected by a new network of 29 seismograph stations have shown, however, that its crustal thickness is only some 45 km. Combined with observations on Pn and Sn phases, this shows that most of the East Anatolian High Plateau is devoid of mantle lithosphere. Areas of no mantle lithosphere is inferred to coincide broadly with the extent of the East Anatolian Accretionary Complex, a subduction-accretion prism of late Cretaceous to earliest Oligocene age. The absence of mantle lithosphere is ascribed to slab breakoff beneath the prism and the widespread vulcanicity to melting its lower levels because of direct contact with hot asthenosphere. The East Anatolian High Plateau is thus supported not by thick crust, but by hot mantle.

INTRODUCTION

The East Anatolian High Plateau [*Figure 1*] is one of the regions of extensive (150,000 km^2) high elevation (average ~ 2km) along the Alpine-Himalayan mountain

system [*Şengör and Kidd*, 1979]. It has attained this height since the Serravallian (*~13 Ma*)[1] following the terminal collision of Arabia with Eurasia [*Şengör and Kidd, 1979; Dewey et al., 1986*]. This has long been ascribed to crustal thickening resulting from post-collisional intra-continental convergence and a present crustal thickness of approximately 55 km has been assumed on the basis of this ascription [*Şengör, 1980*]. New seismic data collected by a network of 29 stations deployed in Eastern Turkey [*Sandvol et al., 2000*] have been used to deduce crustal thickness values using receiver functions [*Zor et al., this volume*]. They yielded the surprising result of ± 45 km average crustal thickness [*Fig. 2*], which is incompatible with the average elevation even if only a normal-thickness mantle lithosphere lid (defined thermally to be above the peridotite solidus) is assumed. Pn velocities and Sn observations under Eastern Turkey [*Gok et al., 2000, 2003 this volume; Lazki et al., 2003 this volume*] lead to the inference that such a lid probably is not present and that the elevation is a result of mantle temperatures typical at least of those of the asthenosphere. The purpose of this paper is to show that this inference is not only compatible with the Cainozoic geological evolution of the region, but is indeed required by it.

OUTLINE GEOLOGY OF THE EAST ANATOLIAN HIGH PLATEAU

The geology of the East Anatolian High Plateau is best presented in terms of its neotectonic and paleotectonic rock packages and structures. The paleotectonic structures of the plateau occur in three major tectonic units:

(1) The Eastern Rhodope-Pontide arc was an ensialic, south-facing magmatic arc of Albian to Oligocene age. It formed by north dipping subduction under the Eurasian

[1] All numerical ages are according to the Geological Society of America Geological Time Scale, 1998.

continental margin [*Yılmaz et al., 1997*]. An extensive zone of backthrusting brings ophiolitic mélange nappes of Cretaceous age onto its southern margin. These are the hintermost parts of the East Anatolian Accretionary Complex [*Şengör and Yılmaz, 1981, figs. 6E and F*].

(2) East Anatolian Accretionary Complex: Wherever the pre-Adilcevaz Limestone [Aquitanian to Burdigalian: ~20-16 Ma: *Şaroğlu and Yılmaz, 1987*] basement is exposed in the East Anatolian High Plateau, it is seen to consist of late Cretaceous (? younger) ophiolitic mélange and Paleocene to late Oligocene flysch sequences [*Figure 1*]. The mélange occurs in imbricate, mainly north-dipping slices commonly incorporating younger flysches in the imbrications [e.g., *Ketin, 1977; Şengör and Yılmaz, 1981, fig. 4; Tüysüz and Erler, 1995*]. The flysch becomes younger from north to south and also its environment becomes progressively shallower from the Cretaceous to the Oligocene. In the north, the Oligocene is already the unconformable cover on the flysch [e.g., *Tüysüz and Erler, 1995*].

That the East Anatolian accretionary Complex has no continental basement is inferred from the fact that its eastern frame (northwest Iran) had become a part of Laurasia by the early Jurassic while its western frame (the Menderes-Taurus block) was still far away from Laurasia [*Şengör and Natal'in, 1996*].

(3) Bitlis-Pötürge Massif: The East Anatolian Accretionary Complex abuts across the Muş suture a series of highly deformed metamorphic massifs that are allochthonous on late Cretaceous and medial Eocene mélange complexes. *Yılmaz [1993]* has shown that this structure was formed by the collision of the easternmost extremity of the Menderes-

Taurus block, constituted by the Bitlis-Pötürge Massif, with the northern margin of the Arabian Platform in the early Eocene.

The neotectonic episode in eastern Turkey commenced when the Adilcevaz Limestone emerged from the sea and throughout the future high plateau terrestrial sedimentation and volcanicity began. The last marine deposits on the East Anatolian High Plateau are of Serravallian age [*Gelati, 1975*]. The plateau must have thus started its rise at the latest around 11 Ma ago. The oldest volcanics of the Plateau are also of the same age, but widespread vulcanicity did not commence until 6 to 7 Ma ago [*Pearce et al., 1990; Keskin et al., 1998*]. The late Miocene to recent sedimentary rocks of the Plateau are dominantly terrestrial conglomerates and sandstones with shales, marls and subordinate evaporates deposited in individual east-west trending basins bounded by thrust faults [*Şengör et al., 1985; Şaroğlu and Yılmaz, 1987*].

The dominant, active structures of the East Anatolian High Plateau are NE-SW and SE-NW striking strike-slip faults with fewer, mainly E-W striking thrust faults [*Şengör et al., 1985; Dewey et al., 1986;* Bozkurt, 2001; *Örgülü et al., 2003, this volume*]. Fold axial trends in Plio-Pleistocene sedimentary rocks are also dominantly E-W [*Şengör et al., 1985; Dewey et al., 1986; Şaroğlu and Yılmaz, 1987*]. These observations show that the plateau has been shortening N-S during at least the last ±15 Ma, but, at present, this shortening is more actively expressed by strike-slip faulting than by thrusting [*Şengör et al., 1985*].

LITHOSPHERE IN EASTERN TURKEY

Zor et al. 2003 [*this volume*] have estimated crustal thicknesses at 29 stations in Eastern Turkey. We have used their estimates to contour the crustal thickness taking into

account the surface geology and the topography [*Figure 2*]. A comparison of *Figure 1* with *Figure 2* illustrates that areas with crust thicker than 45 km are confined to regions outside the Eastern Anatolian Accretionary Complex. An area of thin crust also extends into the northernmost Arabian Plate coinciding with areas of earlier, Aptian-Albian, rifting [*Şengör, 2001*].

One can speculate on the thickness of the mantle lithosphere in Eastern Turkey on the basis of the following assumptions: (1) by defining the base of the lithosphere as an isothermal surface, (2) taking an approximately steady-state heat flux across the lithosphere, (3) assuming a lack of significant lateral crustal inhomogeneities and (4) taking whole lithosphere isostasy with respect to the mid-ocean ridges. In this paper we assumed ρ_c=2.82 kg/m^3 and ρ_m=3.27 kg/m^3 (as compatible with the geological model) and calculated the thickness of the mantle lithosphere using Lachenbruch and Morgan [*1990*], who tabulated heights of lithospheric columns above sea-level for given thicknesses and densities (as function of temperature) for the crust and mantle lithosphere in each column. For a given density contrast and crustal thickness, we simply deduce the necessary mantle lithosphere thickness for thermal isostasy. The results are shown in Figure 2. (Our topographic elevations have been averaged to areas having radii corresponding with the Fresnel zones relating to the receiver function work).

The average crustal thickness of the Eastern Anatolian High Plateau is difficult to explain if it has even a normal-thickness (as opposed to collision-thickened), fore-arc (i.e. cold) mantle lithosphere. If that were the case, the plateau should have stood at an elevation of some 1.5 km height, in places even lower [*Lachenbruch and Morgan, 1990*]. This is clearly not the case. In fact, the areas of thinnest crust are highest in average

elevation. A comparison of [*Figure 2*] with the one exhibited in Figure 1 shows that areas of no mantle lithosphere coincides almost perfectly with the extent of the East Anatolian Accretionary Complex.

INTERPRETATION AND CONLUSIONS

If the East Anatolian Accretionary Complex is bereft of a mantle lithosphere the question is why. Its geological evolution yields the answer. *Figure 3* shows sequential cross-sections roughly along the 42° E meridian. In the early Eocene, the Rhodope-Pontide arc was still active and possessed a large subduction-accretion complex perhaps not unlike the one in Makran today. By late Eocene time, the toe of this accretionary complex may, in some points, have touched the northern margin of the Bitlis-Pötürge Massif (which, by that time, had been welded to Arabia). Throughout the Oligocene, the East Anatolian Accretionary Complex was shortened and thickened above an oceanic lithosphere sliding beneath it. This 'hidden subduction' [*Şengör, 1984*] may have created the last, Oligocene intrusions in the Rhodope-Pontide arc and extrusives to its immediate south [*38.5 Ma: Keskin et al., 1998*]. After the East Anatolian Accretionary Complex thickened to normal continental crustal thickness, subduction was arrested and Arabia-Eurasia convergence began to be accommodated by intracontinental convergence and crustal shortening from the Greater Caucasus to northern Arabian Plate in the beginning of the Miocene (±24 Ma ago). It is likely that slab breakoff commenced some 11 Ma ago at a depth of some 200 km if a subduction dip angle of 45º and convergence velocity of 2.5 cm/a was maintained between 24 Ma and 11 Ma. If the subducting lithosphere remained in contact with the bottom of the East Anatolian Accretionary Complex as far north as its backstop, then the breakoff would have occurred at a depth of some 50 km

and 300 km north of the suture. At that time, the locus of initial collision-related volcanism was thus some 75 km south of the Eastern Pontide backstop margin along the 42°E meridian, assuming plateau-wide homogeneous north-south shortening. 11 Ma ago was the time when the first collision-related magmatism commenced about 200 km N of the present-day suture line [*Keskin et al., 1998*] and when the plateau surface entirely cleared out of water. By 8 Ma ago slab breakoff was probably complete, when post-collisional volcanism became plateau-wide by spreading mainly southward. The inferred depths and timing of breakoff are remarkably consistent with model calculations of Davies and von Blanckenburg (1995) concerning factors governing slab breakoff. When the accretionary complex was still underlain by the slab, its top had to remain below the level of the ocean, because the slab was most likely older than 100 Ma [see reconstructions in *Şengör and Yılmaz, 1981, Masse et al., 1993* and *Şengör and Natal'in, 1996*] and the thickness of the accretionary wedge was most likely thinner than 45 km. The falling off of the slab exposed the underbelly of the East Anatolian Accretionary Complex to at least asthenospheric temperatures, which resulted in its widespread partial melting. The late Miocene to present volcanicity of Eastern Turkey exhibiting a complex composition and geochemistry ranging from andesitic-rhyolitic crustal melts to alkalic olivine basalts probably reflect the rise of the asthenosphere, its adiabatic melting and heating up of the overlying crust [*Keskin, 2003, this volume*].

Figure 4 shows the comparison of two E-W topographic profiles, low-pass filtered at 125 km to eliminate any possible elastic effects, of the mantle plume-generated Ethiopian High Plateau [*Şengör, 2001*] with the East Anatolian High Plateau. Their similarity is striking and most likely points to a common cause. We believe that cause to

be a hot, rising asthenosphere beneath a piece of crustal lithosphere bereft of its mantle component.

The model here proposed has important implications for other regions underlain by very large subduction-accretion complexes, such as wide areas of the Altaids in Central Asia [*Şengör and Natal'in, 1996*] or the Songpan-Ganzi system in China [*Şengör, 1984*]. In such regions, widespread A-type granite magmatism and felsic and intermediate volcanism invades the former subduction-accretion complexes shortly after the cessation of subduction. This 'late-to post-orogenic' magmatism is an important step in converting subduction-accretion complexes into continental crust and is thus critical for our understanding of the evolution of the latter.

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FIGURE CAPTIONS

Figure 1. Simplified geological map and tectonic units of Eastern Anatolian High Plateau (from various sources). Neotectonic structures not shown for clarity.

Figure 2. Crustal Thickness contours (2 km intervals) from Zor et al. [*this volume*] and thickness of the mantle lithosphere in km.

Figure 3. Schematic cross-sectional tectonic evolution of the east Anatolian High Plateau from the Eocene to the present.

Figure 4. Comprasion of the topography of Ethiopia with an E-W profile along the 40˚N parallel in Eastern Anatolia. The smooth lines are least squares simplifications of the topography.

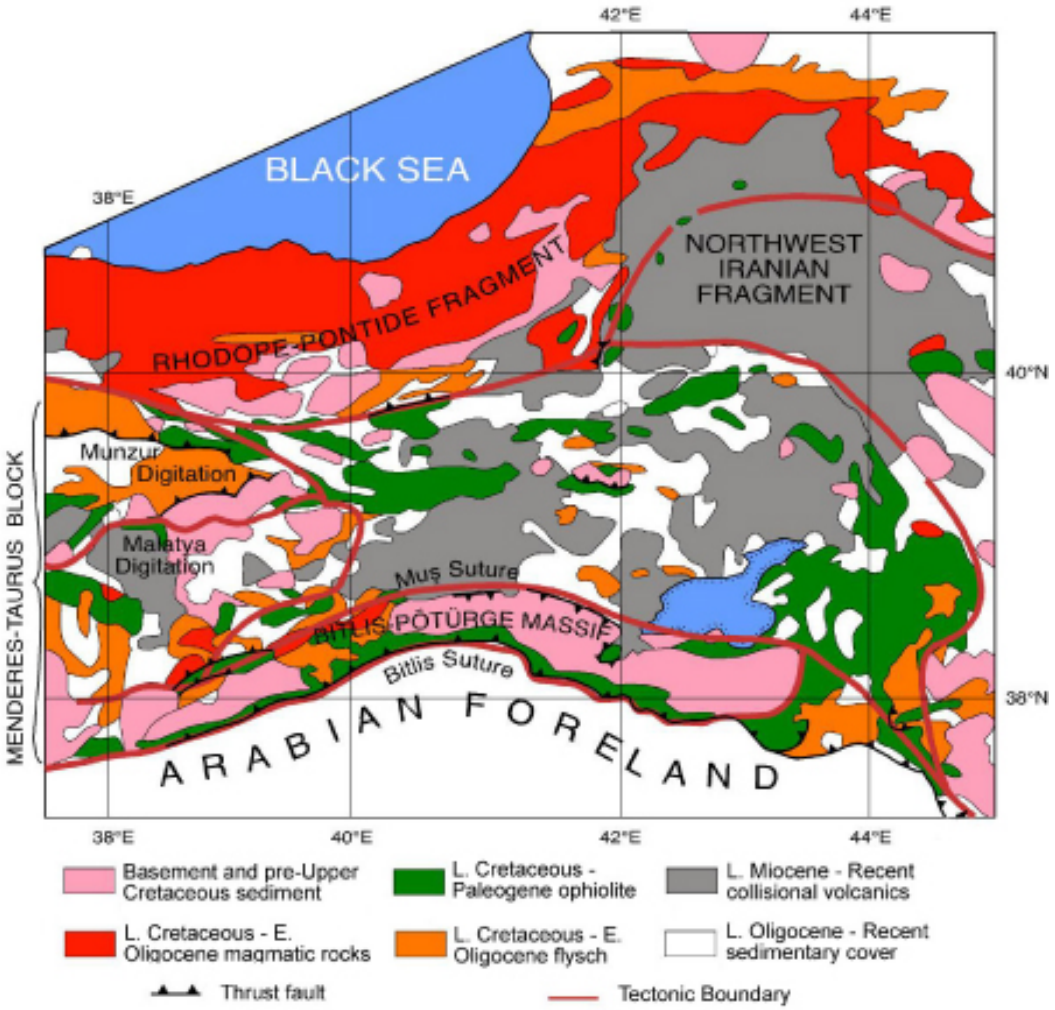

Figure 1

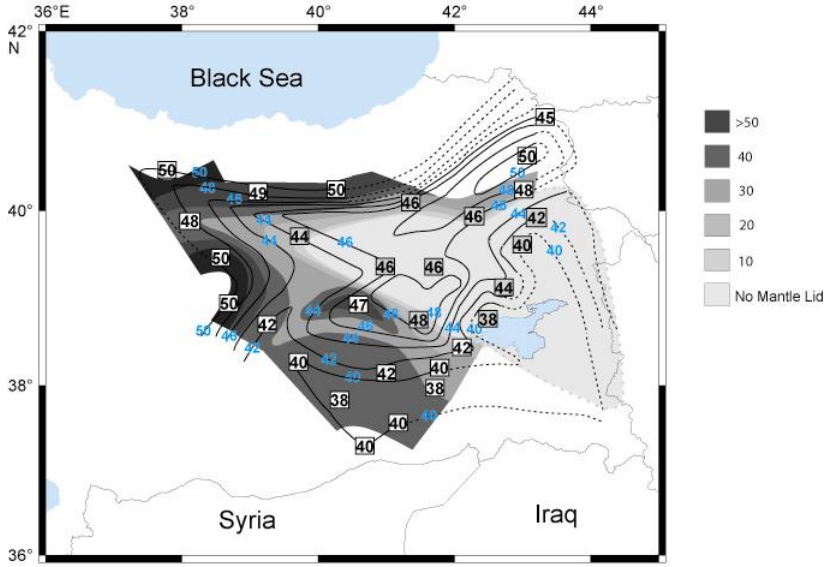

Figure 2

Figure 3

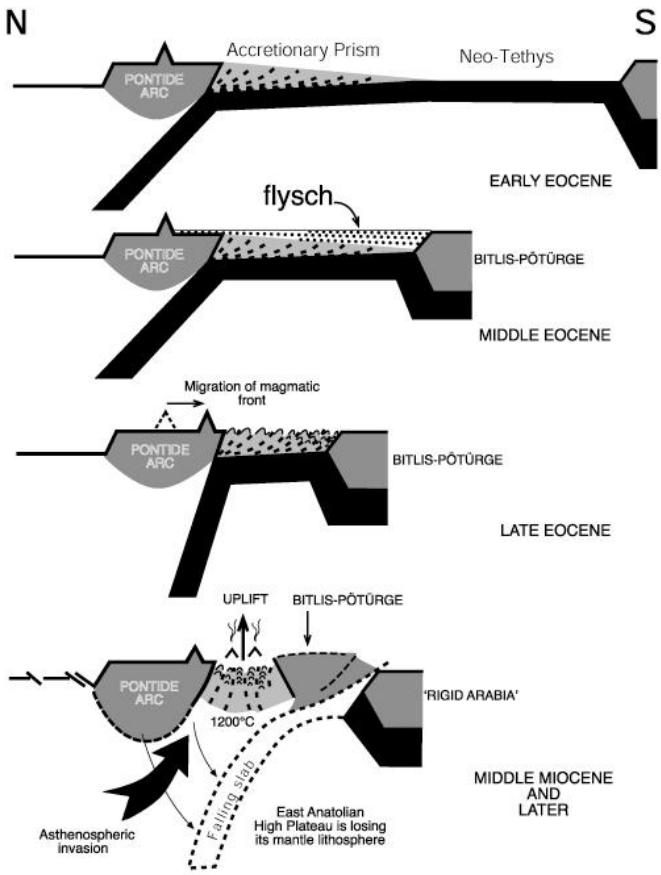

Figure 4