Apatite fission-track data for the Miocene Arabia-Eurasia collision

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ABSTRACT

The collision between the Eurasian and Arabian plates along the 2400-km-long Bitlis-Zagros thrust zone isolated the Mediterranean from the Indian Ocean and has been linked to extension of the Aegean, rifting of the Red Sea, and the formation of the North and East Anatolian fault systems. However, the timing of the collision is poorly constrained, and estimates range from Late Cretaceous to late Miocene. Here, we report the first apatite fission-track (AFT) ages from the Bitlis-Zagros thrust zone. The AFT samples are distributed over the 450 km length of the Bitlis thrust zone in southeast Turkey and include metamorphic rocks and Eocene sandstones. Despite the disparate lithology and large distance, the AFT ages point consistently to exhumation between 18 and 13 Ma. The AFT ages, along with a critical appraisal of regional stratigraphy, indicate that the last oceanic lithosphere between the Arabian and Eurasian plates was consumed by the early Miocene (ca. 20 Ma). The results imply that Aegean extension predated the Arabia-Eurasia collision.

INTRODUCTION

The 2400-km-long Bitlis-Zagros thrust zone between Arabia and Eurasia is one of the largest Tertiary continent-continent collision zones (Fig. 1). The Arabia-Eurasia collision closed the Neotethyan oceanic gateway by isolating the Mediterranean and Indian Oceans; this has been related to mid-Cenozoic global cooling (Allen and Armstrong, 2008). The collision created a wide zone of diffuse deformation on the southern margin of Eurasia (Fig. 1) and has also been linked to the rifting of the Red Sea, extension in the Aegean, and the formation of the North and East Anatolian fault systems (e.g., Jolivet and Faccenna, 2000). Despite its importance, the age of the initial collision between Arabia and Eurasia is poorly constrained, and estimates range from Late Cretaceous (Hall, 1976; Berberian and King, 1981; Alavi, 1994), to late Eocene–Oligocene (35–25 Ma; Jolivet and Faccenna, 2000; Agard et al., 2005; Allen and Armstrong, 2008), to Miocene (Şengör et al., 1985; Dewey et al., 1986; Yılmaz, 1993; Robertson et al., 2007). These estimates are generally based on the stratigraphy and age of deformation of the facing margins of the Arabian and Eurasian plates. A complexity in this context is that both the Arabian and Eurasian margins have been affected by incremental deformation through time. An ophiolite nappe was thrust over the Arabian plate during the Late Cretaceous (ca. 85 Ma), leading to deformation and metamorphism of the passive margin (e.g., Ricou, 1971). The remnants of the ophiolite can be traced from Oman to southeast Turkey over a distance in excess of 2500 km. The opposite Eurasian continental margin was deformed by the late Paleocene–early Eocene (ca. 55 Ma) closure of the more northerly Izmir-Ankara-Erzincan Ocean (Fig. 1; Okay and Tüysüz, 1999) and by the opening and closure of a short-lived Eocene basin that developed on the active margin of Eurasia (Maden-Hakkari basins; e.g., Yılmaz, 1993; Robertson et al., 2007). These deformation events make the dating of the Arabia-Eurasia collision solely on the basis of stratigraphy equivocal.

Collision between two large continental plates is a prolonged process, as exemplified by the ongoing India-Asia collision, which started in the Eocene (e.g., Rowley, 1996). Inception of continental collision is defined by the consumption of the oceanic lithosphere between the continents. This is followed by increased subsidence in the lower plate through thrust loading and a change from shelf to foredeep sedimentation in the associated peripheral foreland basin. Foreland basin successions are expected to evolve from marine sandstone-shale to fluvial coarse clastics as the thrust front moves toward the foreland. Increasing continental shortening will lead to an increase in the crustal thickness, uplift, and erosion, which can be dated by thermochronology.

GEOLOGICAL SETTING

At present, deformation associated with the collision between the Eurasia plate and the Arabia plate is distributed over the Turkish-Iranian Plateau (Reilinger et al., 2006; Copley and Jackson, 2006). The Bitlis-Zagros thrust zone forms the southern boundary of this wide region of deformation. It can be subdivided into the east-west–trending Bitlis and northwest-southeast–trending Zagros thrust zones (Fig. 1). Plate reconstructions indicate that convergence between the Arabian and Eurasian plates was more orthogonal along the Bitlis segment compared to the Zagros segment, which was affected by a considerable degree of dextral strike-slip faulting (McQuarrie et al., 2003; Allen et al.,

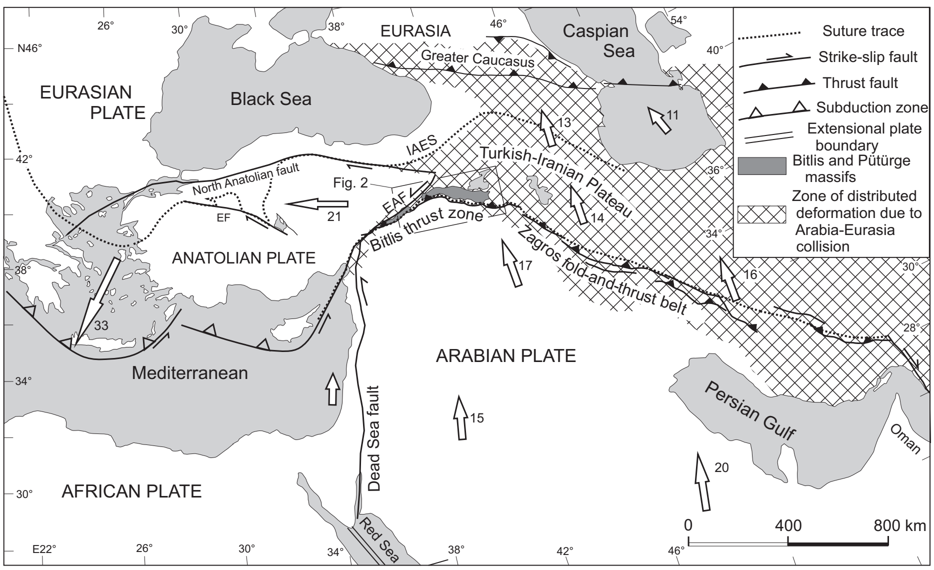

Figure 1. Tectonic map of eastern Mediterranean and Middle East. Arrows and numbers indicate global positioning system (GPS)–derived velocities with respect to Eurasia (modified from Reilinger et al., 2006; Copley and Jackson, 2006). EAF—East Anatolian fault; EF—Eskişehir fault; IAES—İzmir-Ankara-Erzincan suture.

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2004). However, at present, the Zagros segment is characterized by shortening, and the Bitlis segment is characterized by strike-slip faulting (Reilinger et al., 2006).

The 800-km-long Bitlis thrust zone essentially consists of three superimposed tectonic units: (1) the sedimentary succession of the Arabian Platform is overlain by (2) a complex tectonic zone made of ophiolitic mélange and Eocene flysch, and (3) by a metamorphic complex made up of the Bitlis Massif in the east and the Pütürge Massif in the west (Fig. 2; Hall, 1976; Yılmaz, 1993). These massifs consist of south-verging, tectonically imbricated metamorphic rocks forming part of the crustal succession at the leading edge of Eurasia. The metamorphic rocks are subdivided into a Neoproterozoic basement of gneiss, schist, amphibolite and metagranitoid, and a Paleozoic-Mesozoic cover of metaclastic rocks and marble (e.g., Göncüoğlu and Turhan, 1984; Oberhänsli et al., 2009). Remnants of an ophiolite nappe, emplaced during the Late Cretaceous from the north, are locally preserved on top of the Mesozoic metasediments. The Bitlis and Pütürge Massifs underwent metamorphism during the Late Cretaceous under the obducted ophiolite. In the Bitlis Massif, the pressure and temperature during peak metamorphism are estimated at ca. 1 GPa (~33 km depth) and 350–400 °C (Oberhänsli et al., 2009). Ar-Ar and K-Ar white mica ages are ca. 74 Ma in the Bitlis Massif (Göncüoğlu and Turhan, 1984; Oberhänsli et al., 2009), and K/Ar whole-rock ages are 71.2 ± 3.6 Ma in the Pütürge Massif (Hempton, 1985).

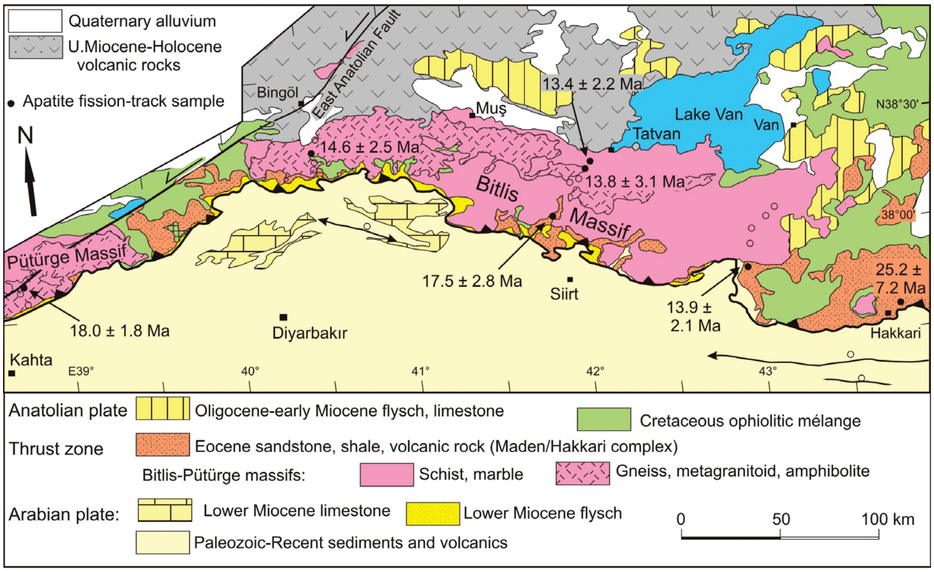

Figure 2. Tectonic map of Bitlis thrust zone with apatite fission-track sample localities and ages.

APATITE FISSION-TRACK DATA

We collected metamorphic rock and Eocene sandstone samples for apatite fission-track (AFT) analysis along four traverses across the Bitlis and Pütürge Massifs and the collision zone (Fig. 2; Table 1). Spot samples were also taken from the Paleozoic sandstones in the collision-induced anticlines on the Arabian plate and from the allochthonous Eocene sandstones around Hakkari. Procedures for sample preparation and analysis are those described in Zattin et al. (2005). Apatite grains from 23 samples were sent for irradiation; however, uranium content in most samples was too low to generate enough tracks for a reliable age. Only seven samples yielded apatites suitable for fission-track analysis. These samples are distributed over a 450-km-wide region and come from two tectonic units (Fig. 2; Table 1): (1) the metagranitoids and the overlying metasediments of the Bitlis and Pütürge Massifs, and (2) Eocene sandstones from the Maden and Hakkari complexes in the collision zone. The Eocene sandstones show a low-grade metamorphism in the Hakkari region but are unmetamorphosed elsewhere (Perinçek, 1990; Oberhänsli et al., 2009).

Despite the lithological diversity and the wide geographic distribution, AFT results range consistently between 18 and 13 Ma (Burdigalian-Langhian) (Table 1; Fig. 3). The Miocene AFT ages from the Eocene sandstones indicate that they were buried below the apatite zone of partial annealing (~110 °C). The AFT ages do not show any correlation with elevation (Table 1). Given the small number of tracks, their lengths could be measured only in three samples, and only one gave a significant number. Data related to this sample were modeled using the HeFTy program (Ehlers et al., 2005). Results show a rapid increase in exhumation at ca. 12 Ma (Fig. 4).

TABLE 1. APATITE FISSION-TRACK DATA FROM THE BITLIS THRUST ZONE

Sample number	Coordinates (UTM)	Elevation (m)	Rock type	No. of crystals	Spontaneous		Induced		$P(\chi)^2$	Dosimeter		Age (Ma ± 1σ)	Mean confined track length (μm)	Std. dev.	No. of tracks measured
					ρ_s	N_s	ρ_i	N_i		ρ_d	N_d				
TU136	38S0251160 4260508	1642	Metasandstone	20	0.72	40	0.89	496	100.0	0.90	4293	13.4 ± 2.2	13.98 ± 0.21	0.92	19
TU138	38S0241967 4249698	1285	Gneiss	16	0.46	22	0.55	264	100.0	0.90	4281	13.8 ± 3.1	–	–	–
TU140	37S0753971 4234870	871	Eocene sandstone	4	5.14	43	4.84	405	91.1	0.90	4256	17.5 ± 2.8	–	–	–
TU145	37S0630748 4277901	1175	Metagranite	20	0.55	38	0.62	425	82.5	0.89	4219	14.6 ± 2.5	–	–	–
TU149	37S0476619 4240707	1395	Gneiss	20	1.60	112	1.44	1006	87.0	0.88	4181	18.0 ± 1.8	13.00 ± 0.23	1.99	72
TU155	38S0321648 4195176	1607	Eocene sandstone	20	0.88	53	1.18	711	65.1	1.01	4818	13.9 ± 2.1	14.51 ± 0.29	1.41	24
TU159	38S0396240 4162747	1342	Eocene sandstone	6	0.53	14	0.39	102	75.4	1.00	4771	25.2 ± 7.2	–	–	–

Note: Central ages were calculated using dosimeter glass CN5 and ζ-CN5 = 367.45 ± 4.35 (analyst MZ). ρ_s—spontaneous track densities (×10^5 cm^{-2}) measured in internal mineral surfaces; N_s—total number of spontaneous tracks; ρ_i and ρ_d—induced and dosimeter track densities (×10^6 cm^{-2}) on external mica detectors (g = 0.5); N_i and N_d—total numbers of tracks; $P(\chi^2)$—probability of obtaining χ^2 value for degrees of freedom (where ν = number of crystals – 1); a probability >5% is indicative of a homogeneous population. Samples with a probability <5% were analyzed with the binomial peak-fitting method.

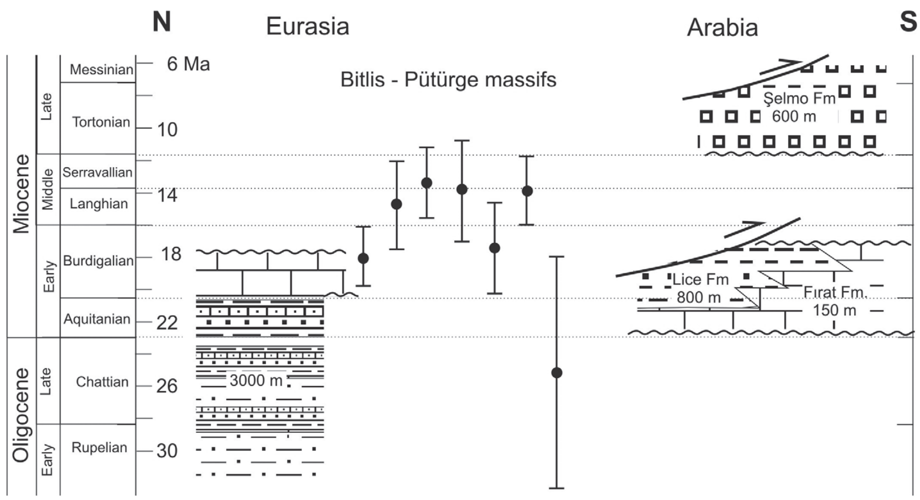

Figure 3. Apatite fission-track ages and regional stratigraphy of Bitlis thrust zone. Geological time scale is from Gradstein et al. (2004).

REGIONAL STRATIGRAPHY

Tectonostratigraphic analysis of sedimentary successions along the facing converging margins of the Arabian and Eurasian plates can provide additional constraints on the timing of collision (cf. Rowley, 1996). Along the northern margin of the Arabian plate in southeast Anatolia, widespread shallow-marine carbonate deposition during the Eocene was followed by an Oligocene regression, and Oligocene deposits are missing over much of the region (Duran et al., 1988; Perinçek et al., 1993). A marine transgression at the beginning of the Miocene led to the deposition of Lower Miocene (Aquitanian-Burdigalian) reefal carbonates, which crop out close to the Bitlis thrust zone (Fig. 2). Such carbonates pass upward and northward to siliciclastic marine turbidites, again of Lower Miocene age (Fig. 3; Duran et al., 1988; Yılmaz and Duran, 1997). This Lice Formation, which is 800 m thick, was deposited in a peripheral foreland basin during its phase of underfilling and is at present partly overthrust by the Bitlis Massif and the Eocene series (Fig. 2). It represents the last marine sequence on the northern margin of the Arabian plate and is followed by the deposition of continental conglomerate, sandstone, and evaporites of late Miocene–Pliocene age.

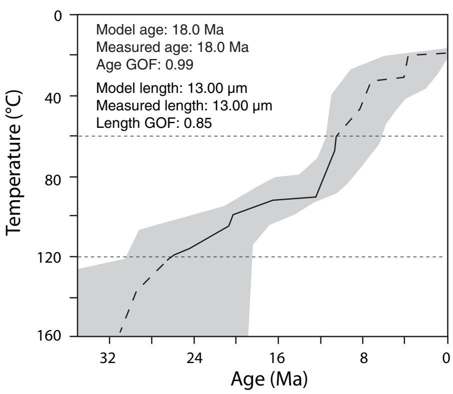

Figure 4. Time-temperature paths for sample TU149 obtained from inverse modeling. Predicted apatite fission-track (AFT) data were calculated according to annealing model of Ketcham et al. (1999) and c-axis projection of Donelick et al. (1999). Shaded areas mark envelopes of statistically acceptable fit, and thick line corresponds to most probable thermal history. Goodness of fit (GOF) gives an indication about fit between observed and predicted data (values close to 1 are best).

North of the suture on the Eurasian plate, a thick succession of Oligocene turbidites crops out north of Muş; these are overlain by shallow marine sandstone, shale, and limestone of early Miocene (Aquitanian-Burdigalian) age (Fig. 2; Sancay et al., 2006). The Oligocene-Miocene succession was deposited in a retroarc foreland basin associated with northward subduction of the Arabian plate. Lower Miocene (Lower Burdigalian) shallow-marine limestones also crop out widely west of Lake Van (Özcan and Less, 2009), where they are unconformably overlain by Pliocene to Holocene volcanic rocks. Widespread volcanism in eastern Turkey, associated with postcollisional slab breakup or mantle delamination, started in the late Miocene (ca. 11 Ma) in the north and migrated southward, reaching the region of Lake Van in the latest Miocene to Pliocene (6–3 Ma; Keskin, 2007).

DISCUSSION

The early to mid-Miocene AFT ages are interpreted as reflecting exhumation along the Bitlis thrust zone due to Arabia-Eurasia collision. This is based on the coherent age spectrum from across the 450-km-long Bitlis thrust zone, similar AFT ages from the Neoproterozoic basement and the Eocene sandstones, and confined track lengths that point to fast rock cooling at ca. 12 Ma. This inference is supported by the regional stratigraphy, which shows increased subsidence in the peripheral foreland basin over the Arabian plate in the early Miocene (Aquitanian-Burdigalian). The Eurasia-derived detritus within the peripheral foreland basin fill indicates that the oceanic lithosphere between the Arabian and Eurasian plates was consumed by the early Miocene (ca. 20 Ma). The stratigraphic record indicates that Eastern Anatolia became a land area only after the early Burdigalian. The first extensive mammal exchange between Arabia and Eurasia was during the late Burdigalian (ca. 18 Ma; Rögl, 1999; Harzhauser et al., 2007), and the Black Sea–Caspian region became isolated from the global marine environment in the Serravallian (ca. 13 Ma) as the result of uplift of the Turkish-Iranian Plateau (e.g., Van Couvering and Miller, 1971).

Thermal paths obtained from modeling show a cooling rate of ~16 °C/m.y. during the period of highest exhumation between 12 and 8 Ma. Taking a geothermal gradient of 25–30 °C/km, based on heat flow (Tezcan, 1995) and the depth of the Curie point in Eastern Anatolia (Aydın et al., 2005), this cooling rate translates into an exhumation rate of 0.5–0.6 km/m.y. Since the Miocene, the average exhumation rate in the Bitlis thrust zone has been 0.1–0.2 km/m.y., which can be compared with the much faster exhumation rates (>1.5 km/m.y.) in the Himalaya (e.g., Blythe et al., 2007).

Present crustal movements along the Bitlis thrust zone change from predominantly strike-slip in the west to orthogonal shortening in the east (Reilinger et al., 2006). However, the AFT ages do not show any systematic change across the Bitlis thrust zone, implying that the present crustal movements did not lead to significant differential exhumation.

CONCLUSIONS

Interpretation of the new apatite fission-track data and appraisal of the regional stratigraphy indicate that, following the elimination of the last oceanic lithosphere, the Arabia and Eurasia plates started to collide along the Bitlis thrust zone in the early Miocene (ca. 20 Ma); the stratigraphic record indicates that the thrust zone was above sea level by the end of the early Miocene (ca. 16 Ma). AFT data do not show any systematic change along the 450-km-long segment of the Bitlis thrust zone, but they do indicate a period of increased exhumation in the mid- to late Miocene. Exhumation of the Bitlis thrust zone since the early Miocene has been modest (3–4 km), with average exhumation rates of 0.2–0.3 km/m.y.

The early to mid-Miocene Arabia-Eurasia collision supports a temporal link between collision and the formation of the late Miocene

North Anatolian fault (e.g., Şengör et al., 2005). However, extension in the Aegean domain, related to the slab retreat, dates back to the Oligocene (e.g., Jolivet and Faccenna, 2000). This and the recent recognition of major Oligocene dextral strike-slip faults in Anatolia (Zattin et al., 2005; Okay et al., 2008) suggest that the westward translation of Anatolia started in the Oligocene and predated the Miocene Arabia-Eurasia collision.

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