Thermal histories from the central Alborz Mountains, northern Iran: Implications for the spatial and temporal distribution of deformation in northern Iran

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ABSTRACT

We integrate new and existing thermochronological, geochronological, and geologic data from the western and central Alborz Mountains of Iran to better constrain the late Cenozoic tectonic evolution of northern Iran in the context of the Arabia-Eurasia collision. New data are presented for two granitic plutons north of the Alborz Range crest. Additional new apatite (U-Th)-He data are also presented for volcanic, intrusive, and detrital apatite grains from two transects south of the range crest. Our most definitive results include zircon and apatite (U-Th)-He and limited K-feldspar $^{40}Ar/^{39}Ar$ thermal history data from the Cretaceous (ca. 98 Ma) Nusha pluton that reveal that the Alborz basement underwent generally slow denudation (~0.1 km/m.y.) as late as 12 Ma with more accelerated exhumation (~0.45 km/m.y.) that likely began shortly after 12 Ma. The Lahijan pluton, a late Neoproterozoic–Cambrian basement exposure near the Caspian shore, records apatite (U-Th)-He closure at 17–13 Ma. Additional (U-Th)-He results from detrital apatites sampled along two separate horizontal transects all consistently yielded latest Miocene to Pliocene apparent ages that imply that even supracrustal cover rocks within the Alborz have undergone significant, regionally extensive exhumation. Overall, our data are consistent with ~5 km of regionally extensive denudation since ca. 12 Ma. The onset of rapid exhumation in the Alborz at ca. 12 Ma appears to be consistent with other timing estimates that place the onset of the Arabia-Eurasia collision between 14 and 10 Ma.

Keywords: exhumation, (U-Th)-He, thermochronology, collision, Alborz Mountains, Iran.

INTRODUCTION

Iran encompasses a large part of the Arabia-Eurasia collision zone (Fig. 1) and is therefore a key area for studying upper-crustal deformation related to young continent-continent collision. However, the spatial and temporal distribution of deformation resulting from the Arabia-Eurasia collision is poorly constrained, and the timing of initial continent-continent collision remains controversial.

Temporal estimates for the onset of the Arabia-Eurasia continent-continent collision include Late Cretaceous (Haynes and McQuillan, 1974; Stocklin, 1974b; Berberian and Berberian, 1981; Alavi, 1994), Eocene (Hempton, 1987), Eocene–Oligocene (Hooper et al., 1994), Oligocene (Yilmaz, 1993), Oligocene–Miocene (Berberian et al., 1982), early to middle Miocene (Robertson, 2000), middle Miocene (Dewey and Şengör, 1979; Şengör and Kidd, 1979), middle to late Miocene (Homke et al., 2004), late Miocene (McQuarrie et al., 2003; Stoneley, 1981), and Pliocene (Philip et al., 1989).

The apparent uncertainty regarding the initiation of collision is compounded by the idea that the Arabia-Eurasia collision underwent a tectonic reorganization event at 5 ± 2 Ma (Wells, 1969; Westaway and Arger, 1994; Axen et al., 2001). This is based on the coincident timing (5 ± 2 Ma) of rapid uplift of the Alborz Ranges, rapid subsidence of the south Caspian Basin, reorganization of the Dead Sea transform fault, Red Sea oceanic spreading, the initiation of the North and East Anatolian faults, and coarse molasse sedimentation in the Zagros belt. Axen at al. (2001) tentatively attribute tectonic reorganization in the Middle East to "choking" of the Neotethyan subduction zone by Arabian continental lithosphere.

Allen et al. (2004) follow Robertson (2000) in placing the onset of Arabia-Eurasia collision at 16–23 Ma and suggest that, once thick crust built up in the Turkish-Iranian Plateau and Greater Caucasus (by 5 ± 2 Ma), convergence shifted to less elevated regions like the Zagros simple folded zone, south Caspian region (including the Alborz), and the foothills of the greater Caucasus, accompanied by westward

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Figure 1. Shaded relief map showing the Arabia-Eurasia collision zone. Iran's border is shown by a fine black line. Labels refer to major physiographic and tectonic units. ZMFF—Zagros Mountain front flexure. Box over the western Alborz Mountains shows the location and coverage of Figure 2.

extrusion of Turkey between the North and East Anatolian faults. This idea implies that deformation in collisional zones builds outward from discrete belts or plateau areas surrounding the suture and propagates into adjacent regions of low relief and crustal thickness. In this model, tectonic reorganization of the Middle East at 5 ± 2 Ma merely represents a shift in the locus of deformation outward over time. The idea that widespread tectonic reorganization is caused by choking of the subduction zone by the subduction of thicker continental crust (e.g., Axen et al., 2001) implies that the deformation need not progress outward over time but begins rapidly throughout a broader region as lithospheric heterogeneities respond to the collision.

We can begin to distinguish between these ideas by looking at the timing of deformation in the Alborz Mountains (Fig. 1), which lie 200 to 600 km northeast of the contact between Eurasia and Arabia. The goal is to determine if the present Alborz topography developed in response to the collision or developed earlier or later.

We present tectonic interpretations on the basis of thermochronological and geochronological data from the west-central Alborz Mountains (Fig. 2), which form parts of the northern margin of the Arabia-Eurasia collision zone and the northern boundary of the eastern Turkish-Iranian Plateau (Fig. 1). The data (1) provide tentative temporal constraints for the onset of late Cenozoic orogenesis at a point on the northern edge of the Arabia-Eurasia collision zone, (2) resolve two precollisional cooling events presumably related to deformation during latest Cretaceous–Paleocene and Eocene time, (3) provide a crystallization age for a large granitic intrusion in the western Alborz (Nusha pluton), and (4) provide limits on the migration of deformation within the western Alborz. This paper does not deal with the pre-Cretaceous thermal or tectonic evolution of the Alborz.

GEOTECTONIC SETTING

Paleogeography and Stratigraphy of the Central Alborz

Pre-Permian strata of central and northern Iran indicate platform conditions throughout an undivided Iranian-Arabian Gondwanan margin. Rifting of central Iran-Alborz away from Zagros-Arabia presumably started in the Late Permian to Early Triassic. After rifting away from Gondwana, these margin fragments were transported north as the Paleo-tethys Ocean (Hercynian Ocean in Berberian and King, 1981) subducted under the Eurasian margin, colliding with Eurasia in Late Triassic to earliest Jurassic time (Stocklin, 1974a; Berberian and King, 1981) (Fig. 3). The Paleo-tethyan suture is generally believed to lie north of the Alborz (Stocklin, 1974b; Berberian and King, 1981; Berberian, 1983). It should be noted, however, that the nature of this suturing event and its position in time and space are poorly understood and remain a fundamental research question in the region. The Lower to Middle Jurassic Shemshak Formation (mainly carbonaceous shale, siltstone, and sandstone) overlies older units in the Alborz along an angular unconformity and is conformably overlain by fossiliferous Jurassic marl and limestone (Assereto, 1966).

Cretaceous carbonates and arc volcanic and older rocks in the south-central Alborz are folded and overlain above an angular unconformity by the Paleocene–Eocene conglomerates (Fajan Formation) and the Eocene Ziarat Formation (nummulitic limestone), which throughout northern Iran marks the base of the Eocene Karaj Formation (Stocklin and Setudehnia, 1977). This Cretaceous to Eocene succession represents the transition from a period of tectonic quiescence during Jurassic and Cretaceous time to a period of regional compression in latest Cretaceous to Paleocene time (Stocklin, 1974b; Berberian and King, 1981) (Fig. 3).

The Karaj Formation consists of calc-alkaline, volcanic, and volcaniclastic rocks and shales (Dedual, 1967; Annells et al., 1975a; Annells et al., 1977; Berberian and Berberian, 1981; Vahdati Daneshmand, 1991) deposited in an arc and backarc setting, probably during extension (Berberian, 1983; Hassanzadeh et al., 2002; Allen et al., 2003a). The Karaj Formation is unconformably overlain by shallow marine rocks, basalt, and conglomerate of earliest Oligocene age. The marine rocks include limestone correlative with the Oligocene–Miocene Qom Formation of Central Iran and are interstratified with 33 Ma basalt. This Oligocene–Miocene succession and the Karaj Formation are unconformably overlain by a >1-km-thick sequence of conglomeratic growth strata (red beds) of probable middle to late Miocene age (Guest, 2004). The early Oligocene sequence probably records extension-related subsidence and volcanism (Hassanzadeh et al., 2002), whereas the Miocene red bed growth strata are attributed to collision-related contraction in the west-central Alborz (Guest, 2004).

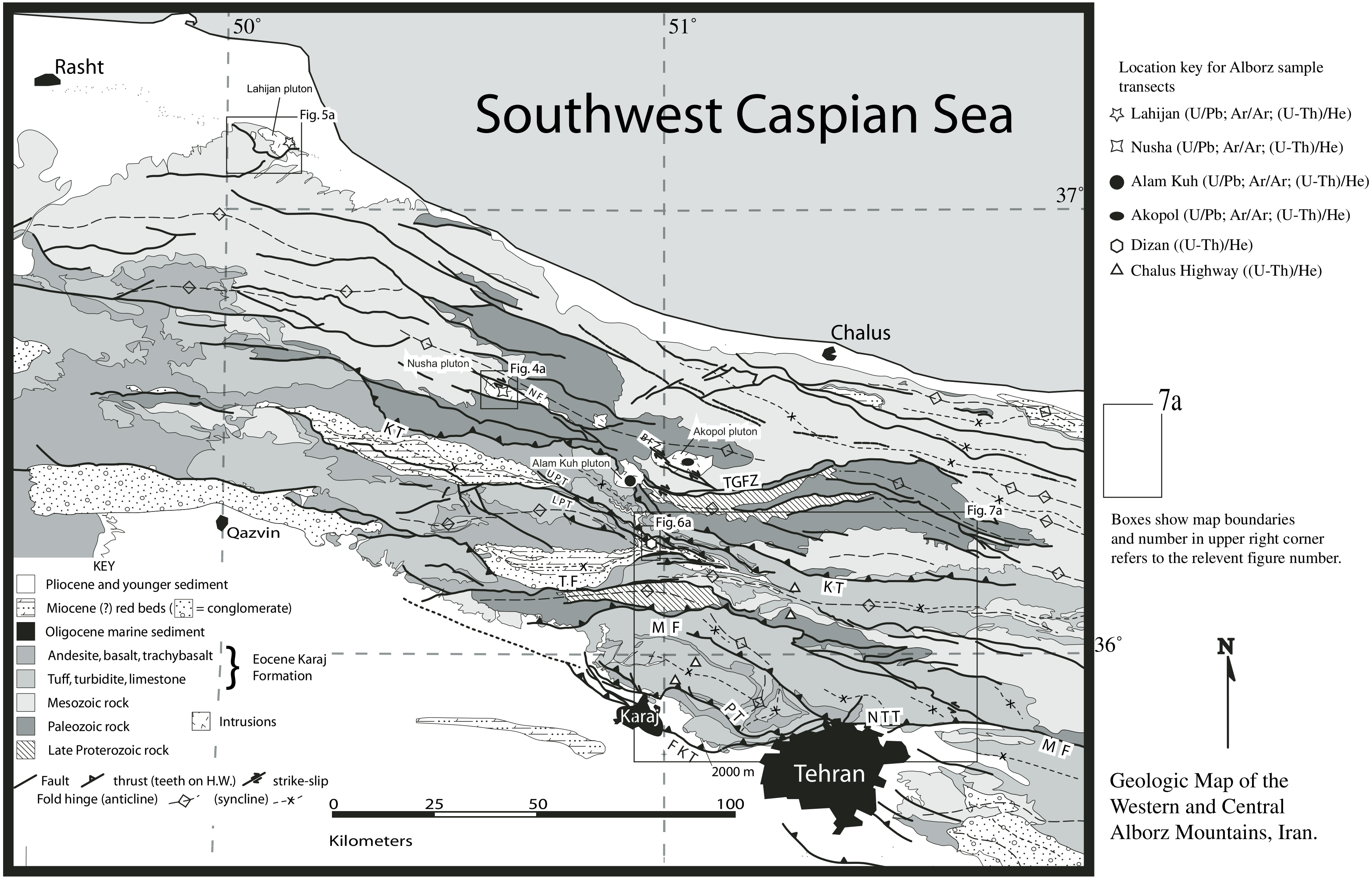


Figure 2. Generalized geologic map of the west-central Alborz Mountains, compiled from our field mapping and the Iranian Geological Survey (IGS) 1:250,000-scale sheets (Annells et al., 1975b; Vahdati Daneshmand, 1991). Dashed boxes show locations and coverage of figures presented in this paper, and the symbols within each box indicate the locations of thermochronometry sample transects. NF—Nusha fault; BFZ—Barir fault zone; KT—Kandavan thrust; TGFZ—Tang-e-Galu fault zone; UPT—upper Parachan thrust; LPT—lower Parachan thrust; TF—Taleghan fault; MF—Mosha fault; NTT—north Tehran thrust; FKT—Fahrahzad-Karaj thrust; PT—Purkan thrust.

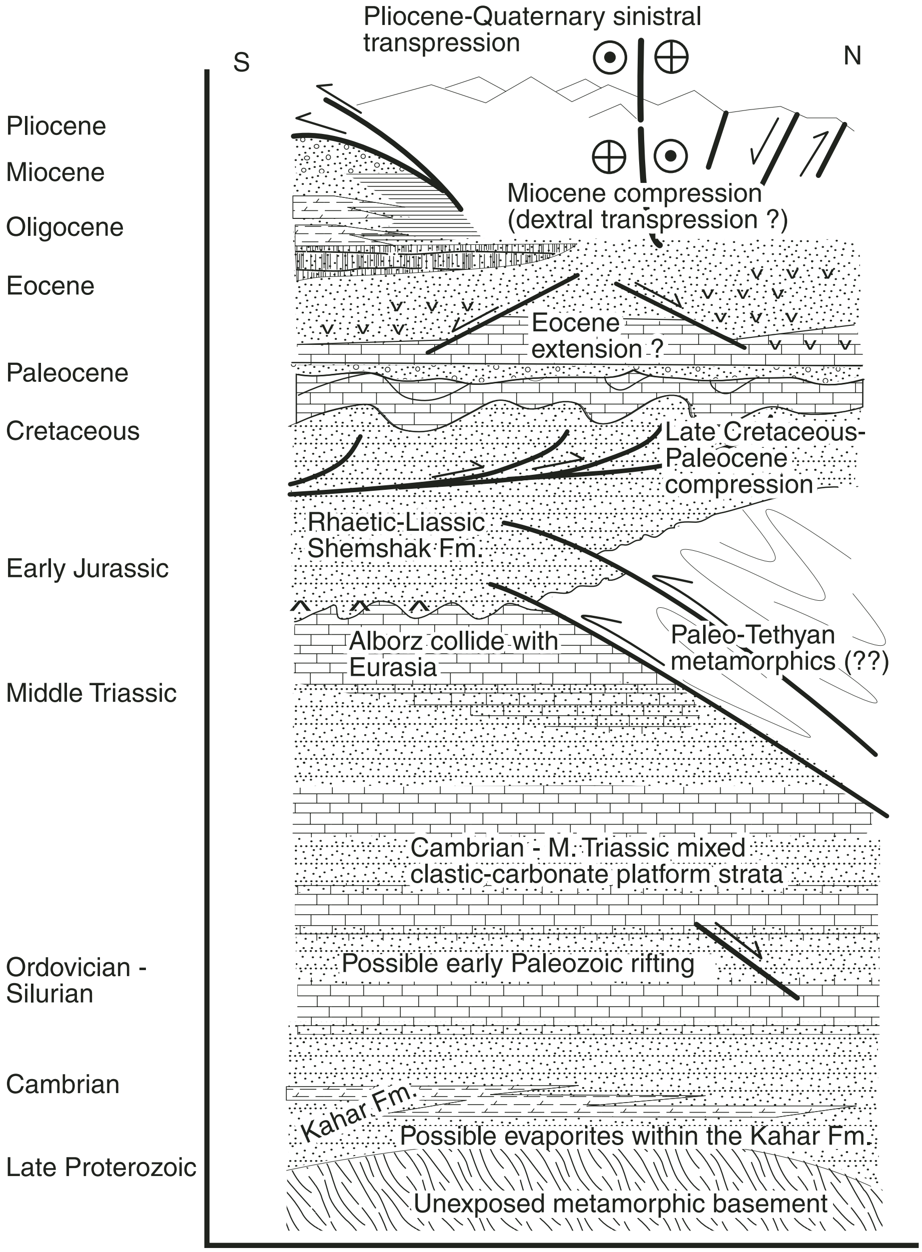


Figure 3. Simplified tectonostratigraphy of the Alborz Mountains, modified from Allen et al. (2003).

Pliocene and younger gravels locally capped by Pliocene and Pleistocene andesite lava flows unconformably overlie folded and faulted Oligocene–Miocene rocks in and along the flanks of the Alborz (Annells et al., 1975a). Outcrops of deformed Pliocene–Pleistocene conglomerate overlain in angular unconformity by gently tilted Quaternary gravels occur in the southern foothills of the range (Annells et al., 1975a; Alavi, 1996). The Quaternary gravels are locally folded and overturned where they are in thrust contact with Eocene or older bedrock along the southern range flank. Pleistocene and younger rocks along the northern range margin are generally flat lying and undeformed (Stocklin 1974a; Annells et al., 1977; Vahdati Daneshmand, 1991).

Deformation

Faults in the Alborz vary in geometry and slip-sense and are typically discontinuous and anastomosing with no single structure dominating the range in terms of slip magnitude (Fig. 2). Faults and folds are oriented subparallel to the range boundaries, and faults generally dip toward the core of the range (Annells et al., 1975a; Annells et al., 1977; Haghipour et al., 1987; Shahrabi, 1991; Vahdati Daneshmand, 1991; Allen et al., 2003b). Earthquake focal mechanisms show that active deformation is partitioned into sinistral strike-slip and reverse-thrust faulting (Priestley et al., 1990; Jackson et al., 2002). It is clear, however, that earlier Cenozoic deformation also was accommodated by dextral strike-slip and normal faulting (Axen et al., 2001; Guest et al., 2006).

THERMOCHRONOLOGY AND GEOCHRONOLOGY

Sampling Strategy

Most of our analytical efforts centered around the use of apatite and zircon (U-Th)-He methods to constrain the denudation history of the upper crust within the Alborz. Two sampling strategies were employed, depending upon our objectives. In selected instances we collected samples along transects in steep areas that afforded sufficient topographic relief between samples (~100 m) to permit us to apply the apatite and zircon (U-Th)-He systems to constrain denudation histories (e.g., Axen et al., 2001; Stockli et al., 2000). (U-Th)-He ages for apatite and zircon in a vertical profile through the upper crust are expected to decrease as a function of depth within the zone of partial He retention. The nature of the profiles depends upon topography and exhumation rates. When topographic controls are well understood, apatite and zircon (U-Th)-He age results from elevation profiles can be analyzed together with estimates of geothermal gradient to estimate denudation history (Reiners et al., 2002; Stockli et al., 2000; Wolf et al., 1998; Wolf et al., 1996). K-feldspar was collected for complementary thermal history modeling on the basis of $^{40}Ar/^{39}Ar$ multidiffusion domain approach (e.g., Lovera et al., 1989, 1997, 2002). We also carried out reconnaissance-style horizontal transects along river valleys or along ridgelines in an attempt to use the apatite (U-Th)-He system to identify and quantify thermal history discontinuities across geological structures. Samples in horizontal transects were collected at ~1 km intervals.

Analytical Methods

The (U-Th)-He procedures employed to determine cooling ages for apatite and zircon follow the methods described by House et al. (2000) and Takahiro et al. (2003) for apatite and zircon, respectively. All analyses were carried out on single grains. Data tables are presented in the GSA Data Repository (DR1.3).[1] Experimentally derived diffusion parameters suggest that He is not retained in apatite crystals above ~80 °C, is partially retained between ~80 °C and ~40 °C, and entirely retained below ~40 °C (Farley, 2000; Stockli et al., 2000; Wolf et al., 1998; Wolf et al., 1996). Diffusion experiments by Farley (2000) suggest a bulk closure temperature (T_c) for He in apatite of ~65 °C to ~75 °C, assuming monotonic cooling at 10 °C/m.y. The zircon (U-Th)-He system diffusion experiments suggest a minimum T_c for He in zircon of 190 °C (for monotonic cooling at 10 °C/m.y.; Reiners et al., 2002). The CLOSURE program by Brandon (2002), available at www.geology.yale.edu/~brandon/, provides a T_c range of ~150 °C to 201 °C for cooling rates ranging from 0.1 °C/m.y. to 40 °C/m.y., respectively. This program uses Dodson's (1973) approach to calculate values of T_c as a function of cooling rate from effective diffusion radius *(r),* and Arrhenius parameters determined by Reiners et al. (2002).

Limited U-Pb analysis of zircon was carried out using the UCLA Cameca ims-1270 ion-microprobe to determine the crystallization ages of two plutonic bodies within the Alborz Ranges. The analytical procedures employed are described in Lam (2002) and Schmitt et al. (2003). All ages are reported with 2σ standard errors. Additional details and U-Pb data tables and plots are presented in the Data Repository (DR1.1; see footnote 1). K-feldspar $^{40}Ar/^{39}Ar$ thermal history modeling was carried out at UCLA using procedures described by Quidelleur et al. (1997), Harrison et al. (1994), and Lovera et al. (1989, 1997). Additional details, data tables, and relevant graphs are presented in the Data Repository (DR1.2; see footnote 1).

TRANSECT DESCRIPTIONS AND RESULTS

Nusha Transect

A 2000 m vertical transect was collected within the Cretaceous Nusha pluton (Fig. 4A). A northwest-southeast–striking dextral fault cuts the pluton (Fig. 2). Only the western half of the body was examined. The body appears to be compositionally diverse, with dioritic compositions encountered at low elevations and rapakivi granite found at the highest levels. Field observations suggested a gradational contact between the diorite and granite phases. The diorite phase contains plagioclase, clinopyroxene, hornblende, and biotite. The granite phase contains plagioclase, coarse (up to 1.5-cm-long) perthite, altered biotite, and hornblende. Titanite, zircon, and apatite are accessory minerals. Zircons from both dioritic and felsic phases were analyzed by Lam (2002; Data Repository Table DR1.1.1a and b; see footnote 1) and yielded statistically indistinguishable weighted mean ion microprobe $^{206}Pb/^{238}U$ ages (97.4 ± 1.8 Ma; MSWD [mean square of weighted deviates] = 1.0 and 96.9 ± 2.4, and MSWD = 0.4 for the mafic and felsic phases, respectively; see Table DR1.1.1b). These results imply that the Nusha intrusion is either a single pluton or a plutonic complex that formed in a single magmatic event.

Samples collected along the transect included two diorites at ~1450 m and 13 granites between 2200 m and 3500 m elevation (Fig. 4A). Six additional samples between 2000 and 2650 m elevation are from the Shemshak Formation north of the Nusha fault (Fig. 4A). (U-Th)-He results for single zircon grains from five samples, including the highest (20–16–1) and lowest (20–25–2), yielded ages between 12 and 35 Ma, which increased with elevation (Data Repository Table DR1.3.2a [see footnote 1] and Fig. 4B). Interpreting the elevation versus apparent age results in terms of a constant geothermal gradient implies a relatively modest exhumation rate (0.086 + 0.04, –0.012 km/m.y.) between 12 and 35 Ma. This is consistent with a cooling rate of 2.1 °C/m.y., assuming a 25 °C/km geothermal gradient and bulk He closure at 174 °C (Brandon, 2002). A multidiffusion-domain, thermal-history interpretation of $^{40}Ar/^{39}Ar$ step-heating results from a single K-feldspar along the traverse (sample 20–24–1) indicates variable cooling, though generally slow cooling, throughout the latest Cretaceous and Cenozoic and appears consistent with the zircon (U-Th)-He results (Fig. 4C). The K-feldspar results also constrain temperatures to have been below 150 °C by ca. 10 Ma. In contrast, apatite (U-Th)-He results from eight samples along the traverse indicated much greater exhumation rates after 7 Ma. The apatite (U-Th)-He ages increase with elevation from 4 to 7 Ma (Data Repository Table DR 1.3.2b [see footnote 1] and Fig. 4B) and suggest an exhumation rate of 0.5 km/m.y., which corresponds to a cooling rate of 10.3 °C/m.y., assuming a constant 25 °C/km geothermal gradient; Fig. 4B).

Thus, on the basis of both the (U-Th)-He apatite and zircon and K-feldspar $^{40}Ar/^{39}Ar$ thermal history results, it seems clear that an approximate fivefold increase in denudation rate occurred somewhere between ca. 7 Ma (the oldest apatite sample age) and ca. 12 Ma (the youngest zircon sample age). It is possible to further refine the time at which the transition to faster denudation took place if the ~100 °C contrast in bulk He closure between apatite and zircon is taken into account. For example, if the transition occurred as late as 7 Ma, the measured zircon (U-Th)-He ages should have been much older than those observed, given the denudation rate implied by the zircon results. Alternatively, the measured zircon (U-Th)-He ages are consistent with the transition that took place near 12 Ma. While more definitive results could have been obtained if more relief had been accessible in our elevation profile, we tentatively conclude that the onset of more rapid exhumation took place ca. 12 Ma. Finally, the fact that (U-Th)-He ages measured from detrital apatites from the Shemshak sandstone we examined from north of the Nusha fault (sample 20–21–1; 2630 m) were similar to those determined for the plutonic samples (Fig. 4A and C) seems to require that faulting either significantly predated late Cenozoic exhumation and/or that minimal <7 Ma dip-slip displacement occurred along the fault.

Lahijan Transect

The Lahijan Granite (Figs. 2 and 5A) is a medium- to coarse-grained biotite granite containing perthitic feldspars and partially altered biotite. Accessory minerals include zircon, titanite, apatite, and epidote. Annells et al. (1975b) interpreted the contact between the Lahijan granite and the surrounding Jurassic and Cretaceous country rock as intrusive, but our reconnaissance investigation indicates that the contact is faulted. In addition, Annells et al. (1975a) reported "Lahijan" pebbles in nearby Jurassic conglomerates and thus concluded that initial erosion of the pluton had begun by Early to Middle Jurassic time.

We collected eight samples over an elevation range of 500 m from the lowest outcrops to the summit ridge south of Layla Kuh village (Fig. 5A). We conducted U-Pb, $^{40}Ar/^{39}Ar$, and (U-Th)-He analyses on zircons, K-feldspar, and apatite (Data Repository Tables DR1.1.1, DR1.2.2, DR1.3.1a, and DR1.3.1b; see footnote 1). Measured ion probe $^{206}Pb/^{238}U$ dates from sample LJ006 indicate a possible late Neoproterozoic to Cambrian crystallization age for the Lahijan granite (Lam, 2002). The zircon (U-Th)-He results indicate He bulk closure by Middle Jurassic to Early Cretaceous time with average ages of 133 ± 7.98 Ma, 152 ± 9.12 Ma, and 162 ± 9.72 Ma for samples LJ003, LJ006, and LJ008, respectively (Tc = ~175 °C, assuming a cooling rate of 4 °C/m.y.; Brandon,

[1]GSA Data Repository item 2006189, methodology, data tables, and figures for U-Pb, $^{40}Ar/^{39}Ar$, and (U-Th)/He analyses, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

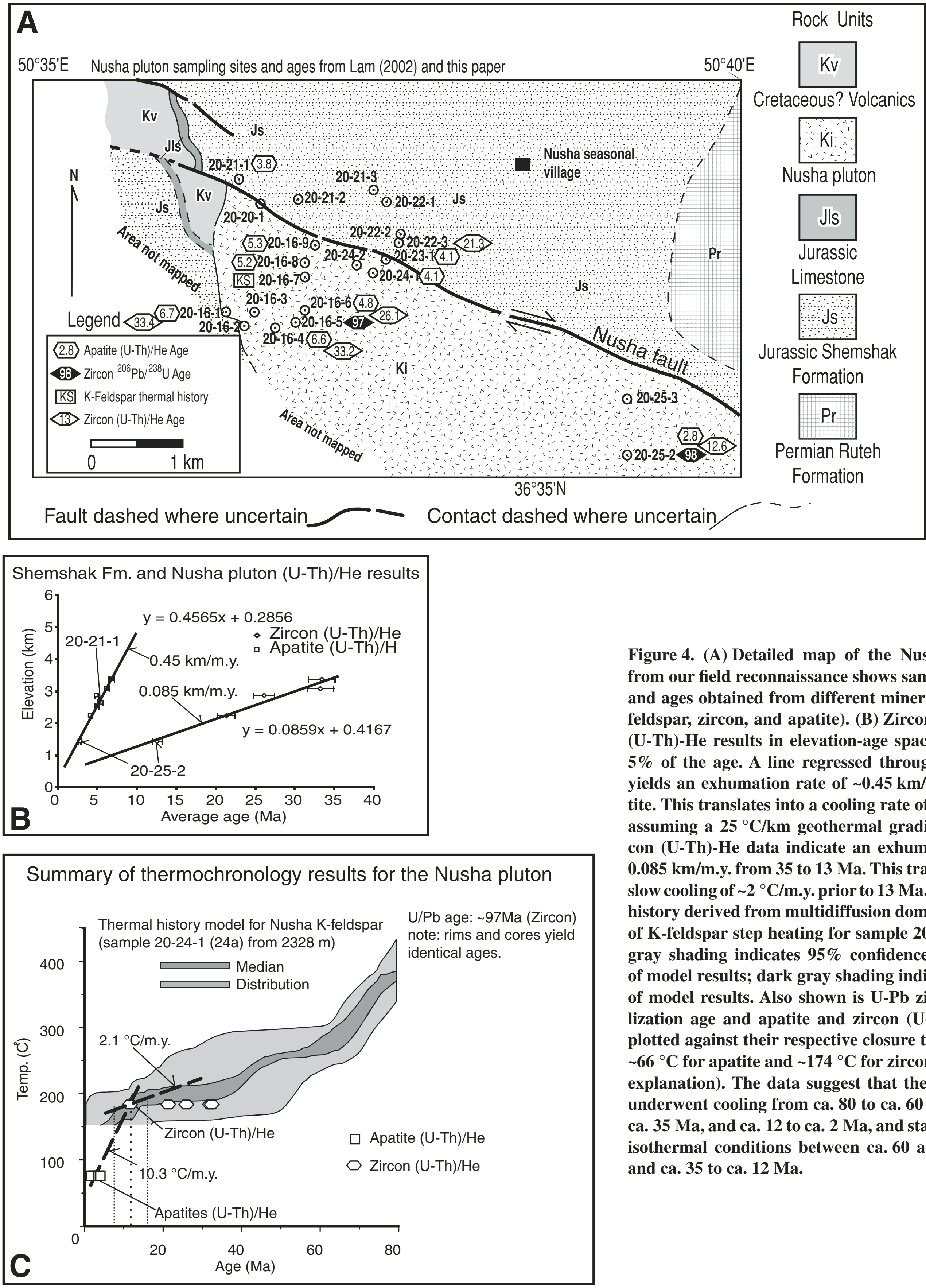
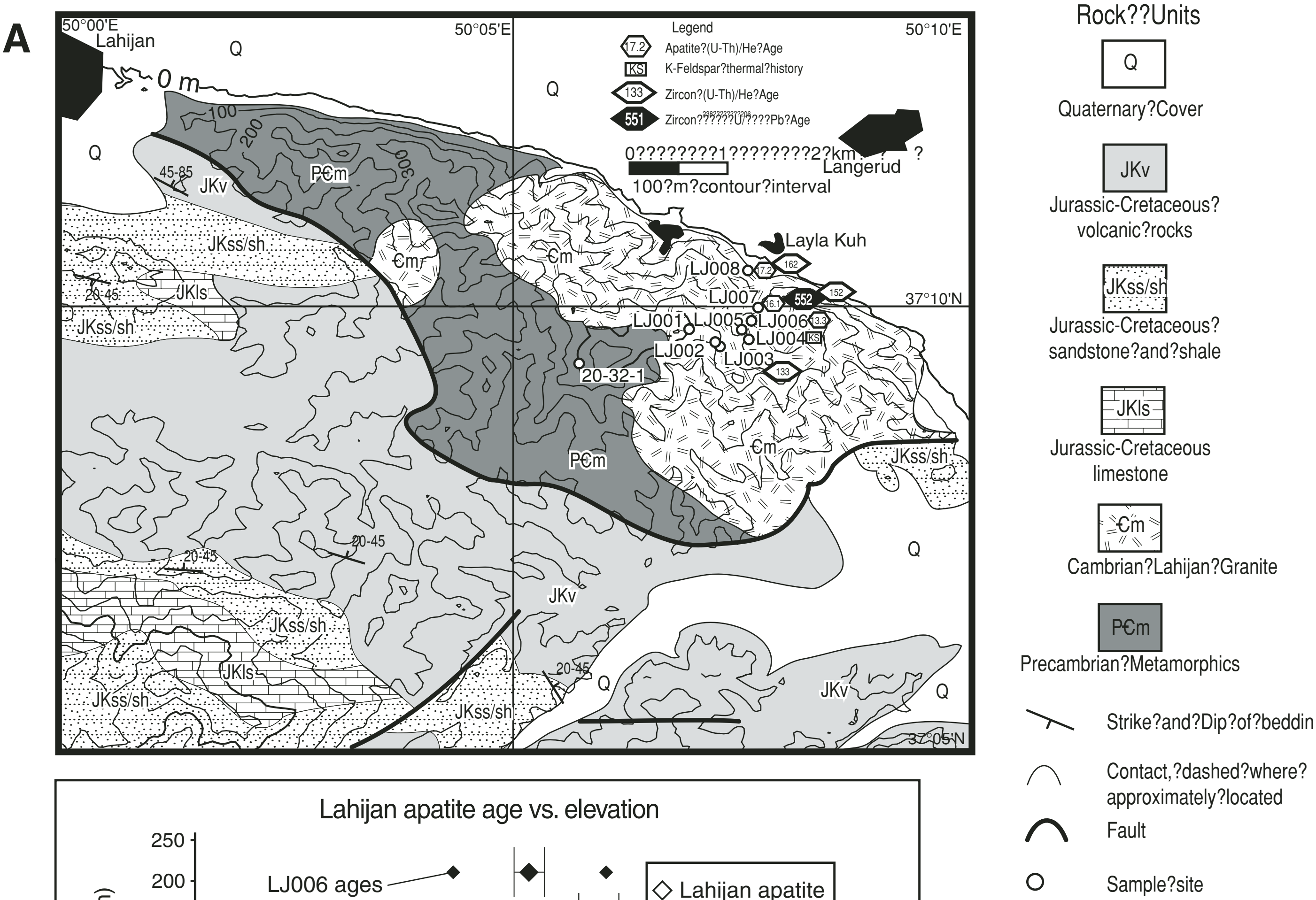


Figure 4. (A) Detailed map of the Nusha intrusion from our field reconnaissance shows sample locations and ages obtained from different mineral phases (K-feldspar, zircon, and apatite). (B) Zircon and apatite (U-Th)-He results in elevation-age space; errors are 5% of the age. A line regressed through these data yields an exhumation rate of ~0.45 km/m.y. for apatite. This translates into a cooling rate of ~10 °C/m.y., assuming a 25 °C/km geothermal gradient. The zircon (U-Th)-He data indicate an exhumation rate of 0.085 km/m.y. from 35 to 13 Ma. This translates into a slow cooling of ~2 °C/m.y. prior to 13 Ma. (C) Thermal history derived from multidiffusion domain modeling of K-feldspar step heating for sample 20–24–1. Light gray shading indicates 95% confidence distribution of model results; dark gray shading indicates median of model results. Also shown is U-Pb zircon crystallization age and apatite and zircon (U-Th)-He ages plotted against their respective closure temperatures, ~66 °C for apatite and ~174 °C for zircon (see text for explanation). The data suggest that the Nusha body underwent cooling from ca. 80 to ca. 60 Ma, ca. 47 to ca. 35 Ma, and ca. 12 to ca. 2 Ma, and stayed at nearly isothermal conditions between ca. 60 and ca. 47 Ma and ca. 35 to ca. 12 Ma.



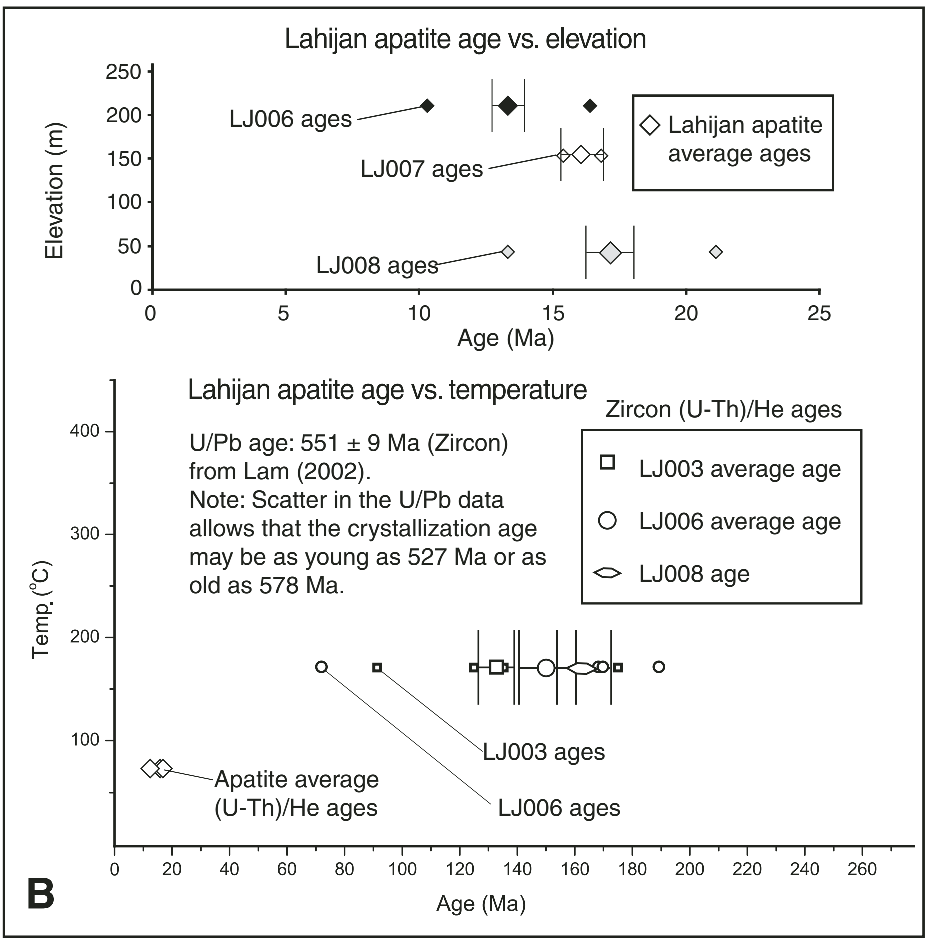


Figure 5. (A) Generalized geologic map of the Lahijan intrusion and surrounding geology, based on our reconnaissance and the Geological Survey of Iran's Qazvin Rasht sheet (1:250,000 scale; Annells et al., 1975b). The map shows sample locations and ages obtained from different mineral phases (zircon and apatite). (B) Upper plot shows apatite (U-Th)-He results plotted in elevation-age space. Lower plot shows apatite and zircon (U-Th)-He results plotted in temperature-time space, assuming 75 °C and 180 °C closure temperature for apatite and zircon, respectively. These data, though somewhat scattered, suggest slow cooling for the Lahijan pluton since Middle Jurassic time.

2002). While these are not in agreement with monotonic thermal histories calculated from the K-feldspar $^{40}Ar/^{39}Ar$ data, the latter are not robust inasmuch as the measured age spectra are poorly fit by the multidiffusion-domain results (Data Repository Fig. DR1.2.1; see footnote 1). The K-feldspar systematics could be explained by transient heating during latest Cretaceous or early Tertiary time, consistent with the presence of Late Cretaceous (Lam, 2002; this paper) and Paleocene plutons (Axen et al. 2001) elsewhere in the range. However, the lack of detailed geochronology, mapping, and structural data from the Lahijan area makes model calculations that assume reheating impractical, as it would be impossible at this stage to determine the extent of, and mechanism for, reheating.

Two apatite grains were dated from each of three samples, producing apatite (U-Th)-He average ages that cluster in the middle Miocene (Data Repository Table DR1.3.2b, see footnote 1; Fig. 5B). These ages indicate that the Lahijan granite cooled through the ~75 °C isotherm during early to middle Miocene time (Fig. 5B). It is unclear what causes the scatter in these data, but it may be explained by slight variations in diffusive properties, long residence times in the apatite partial retention zone, and inaccuracies in grain-size measurements. Slight variations in diffusive properties from grain to grain or inaccuracies in corrections for recoil-induced He loss can lead to large differences in the total amount of He retained after 10 m.y. (Farley et al., 1996).

Dizan Transect

This traverse (Fig. 6A) rises 780 m over a horizontal distance of ~3 km and transects the Parachan thrust system (Data Repository Table DR1.3.3; see footnote 1). The topographically lowest samples we collected were Oligocene–Miocene clastic sedimentary rocks within the footwall of the Parachan thrust system. Two samples of detrital apatite (20-35-1 and 20-36-1) yielded U-Th-He ages of 3.4 ± 0.1 Ma and 4.7 ± 0.3 Ma, respectively (Fig. 6A). Above the lower Parachan thrust (i.e., within the middle plate), a sample of trachyandesite from the Karaj Formation yielded an apatite U-Th-He age of 6.1 ± 0.3 Ma (sample 20–36–3; Fig. 6A). Structurally and topographically higher within the upper plate (i.e., above the upper Parachan thrust) we sampled the Karaj Formation and obtained a (U-Th)-He age from detrital apatite of 4.2 ± 0.2 Ma (Fig. 6A).

The interpretation of these data is strongly limited by the small number of analyzed samples, the fact that the samples are detrital, and an incomplete knowledge of the paleotopography and stratigraphic thicknesses in this region. Nevertheless, the overall similarity in apatite U-Th-He age, coupled with the fact that all measured ages are significantly younger than their sedimentary or volcanic protoliths, strongly implies a shared history during which all of the supracrustal rocks have been exhumed from >3 km depth during late Miocene–Pliocene time. Moreover, the absence of significant age discontinuities across faults of the Parachan thrust system implies that rocks of appreciably different structural levels have not been juxtaposed since the late Miocene–Pliocene.

Chalus Road Transect

Samples on this transect were collected along the Karaj-Chalus Highway, which crosses the Alborz northwest of Tehran (Data Repository Table DR1.3.4, see footnote 1; Fig. 7A). This ~65 km transect starts ~10 km north of Karaj at ~1800 m, crosses the ~4000 m range crest, and continues northeast another ~10 km to a point at 2300 m elevation (Fig. 7A). Samples were collected mainly from volcaniclastic and siliciclastic units of Tertiary, Mesozoic, and Paleozoic age; one sample (KJ012) is from a Tertiary diorite.

Apatite was analyzed by the (U-Th)-He method from seven samples along this transect. The ages shown in Figure 7 represent the average age obtained from two multigrain analyses obtained for each sample (except for sample KJ003 from which we report only one of the ages obtained) (Data Repository Table DR1.3.4, see footnote 1; Fig. 7B). Whereas the interpretation of these results is limited by the level of knowledge of the geology along the transect, the consistency of the apatite (U-Th)-He ages makes this a moot point (Fig. 7A and B). The fact that all of the Karaj Highway samples record cooling between 7 and 4 Ma (0.2–0.4 Ma uncertainty) suggests that the entire south side of the Alborz Range was undergoing significant (<2–3 km) denudation by latest Miocene time.

DISCUSSION

Both new and previously obtained apatite and zircon (U-Th)-He and K-feldspar $^{40}Ar/^{39}Ar$ thermal-history data reveal that both basement and supracrustal rocks underlying the Alborz Range of northern Iran have undergone a relatively coherent late Miocene–Pliocene exhumation history related to the continuing Arabia-Eurasia collision in this region. In the following sections we compile our data with previous results from the Akapol and Alam Kuh plutons (Fig. 2) presented by Axen et al. (2001) to refine previously advanced models for the geologic development of the Alborz region (Fig. 8).

Intrusive Events

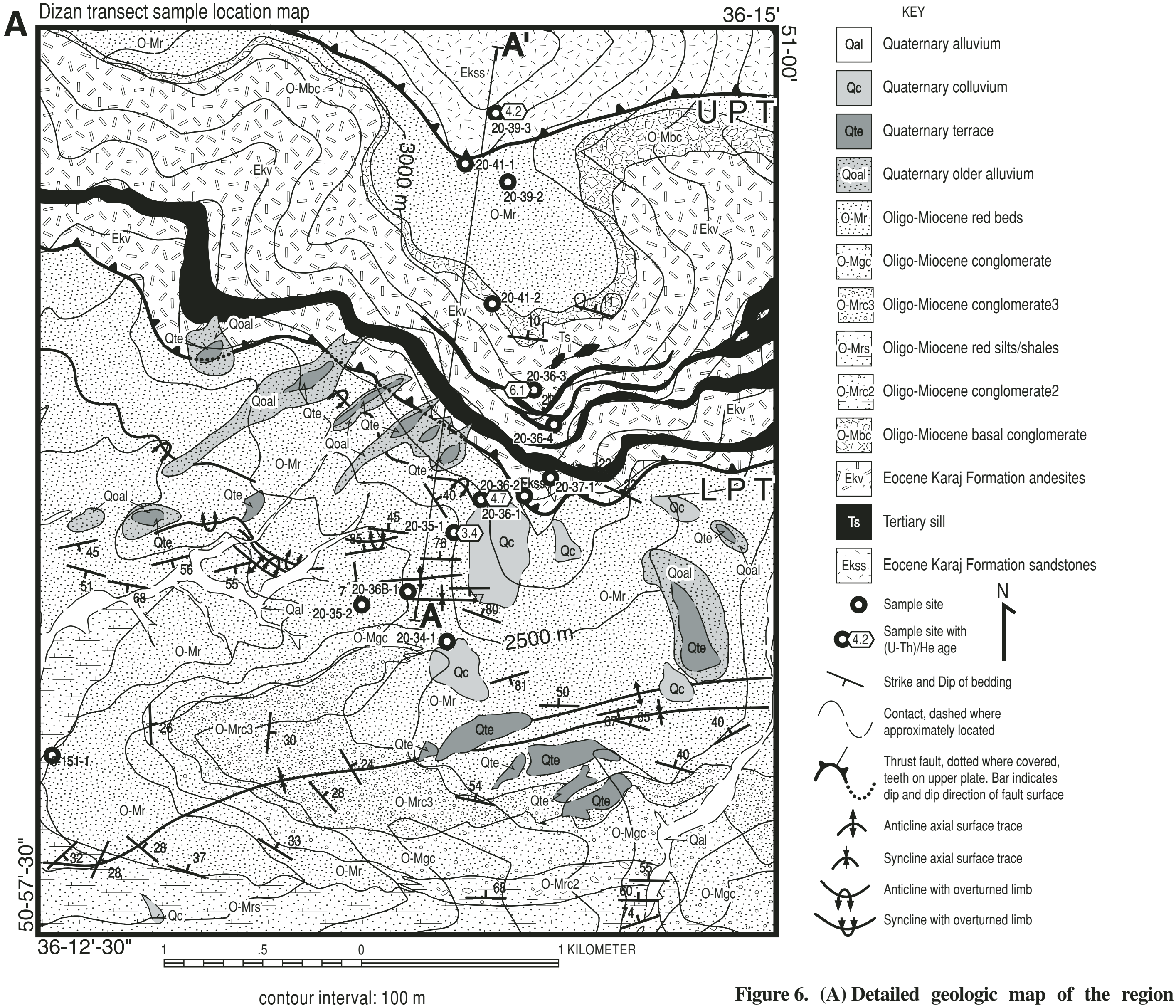
There are no known exposures of crystalline basement rocks in the Alborz. Consequently, the new geochronological results from the Lahijan and Nusha plutons have important implications for our understanding of the tectono-plutonic development of northern Iran. The U-Pb ion probe data for the Lahijan pluton suggest intrusion within the late Neoproterozoic to Cambrian time interval (Lam, 2002).

Intrusive and volcanic rocks of comparable age in central Iran have been attributed to the Peri-Gondwanan (or Proto-Tethyan) orogenic episode (Ramezani and Tucker, 2003). The presence of Peri-Gondwanan intrusions to the south (central Iran) and to the north (Lahijan) of the Alborz Ranges confirms the notion that all lithosphere south of the Caspian was once part of the undivided Gondwana margin (Stocklin, 1974b; Berberian and Berberian, 1981; Ramezani and Tucker, 2003). Thus, the suture between Turan and northern Iran might lie to the north of the Lahijan pluton.

The 97 ± 2 Ma age of the Nusha pluton indicates intrusion during an Aptian to Maastrichtian (121–65 Ma) period of magmatism and carbonate deposition in the western Alborz and Talesh Mountains (Annells et al., 1975a; Clark et al., 1975). If this Alborz-Talesh Cretaceous magmatism was caused by Neotethyan subduction along the present Zagros-Bitlis suture (Berberian and Berberian, 1981; Berberian and King, 1981), then the present width of the Cretaceous arc is ~380 km, measured from arc-related Cretaceous igneous rocks just north of the suture to outcrops in the Alborz and Talesh. This implies either that Cretaceous arc volcanism was spread over an anomalously wide area in Iran, as suggested for Eocene arc volcanism (Berberian and Berberian, 1981), or that this volcanism was concentrated along an initially narrower volcanic arc that was later rifted apart, thereby distributing the arc rocks over a wider region (e.g., Hassanzadeh et al., 2002). The latter idea is consistent with evidence for Eocene extensional subsidence present in the southwestern Alborz, and Eocene backarc or intra-arc rift volcanism in central Iran (Hassanzadeh et al., 2002; Guest, 2004; this study).

Pre-Neogene Tectonism

There is geological evidence for orogenic and taphrogenic events that predate the Arabia-Eurasia collision in Iran (Berberian and King, 1981; Berberian, 1983; Şengör, 1990; Şengör and Natal'in, 1996) (Fig. 3). Our thermal history results appear to correlate with the important post-Jurassic tectonically active and quiescent



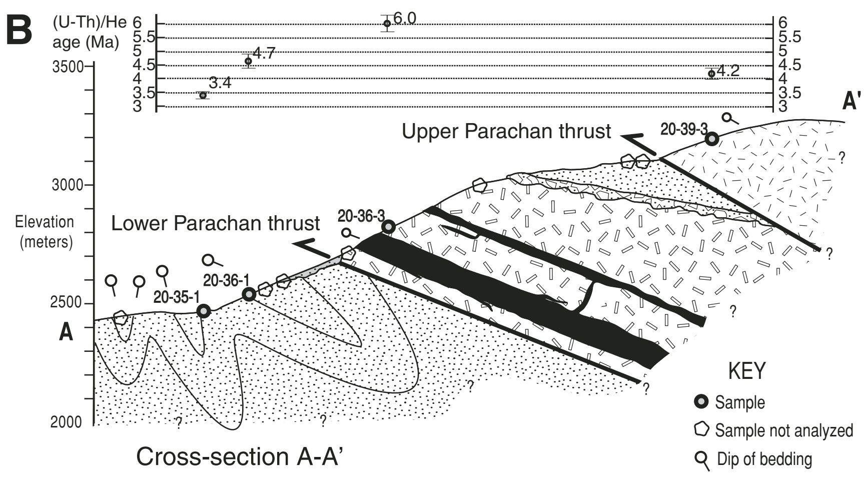


Figure 6. (A) Detailed geologic map of the region around the Dizan (U-Th)-He transect, based on our field mapping. The map shows sample locations relative to surrounding structures and the location of cross-section A–A′. LPT—lower Parachan thrust; UPT—upper Parachan thrust. (B) Cross section A–A′, with sample locations and (U-Th)-He cooling ages plotted above the cross section. The lower three samples cross the lower Parachan thrust and indicate slow cooling from 3.4 to 6.1 Ma. This distribution of ages in highly deformed sediments across a thrust suggests that thrusting and synchronous deformation of the footwall rocks were occurring at temperatures above the closure temperature for apatite (~75 °C) and therefore at a depth >~2 km, assuming a 25 °C/km geotherm and a ~25 °C average surface temperature. The hanging wall of the upper Parachan thrust yields a younger 4.2 Ma age, which suggests the possibility that the upper Parachan thrust remained active after activity on the lower Parachan thrust ceased.

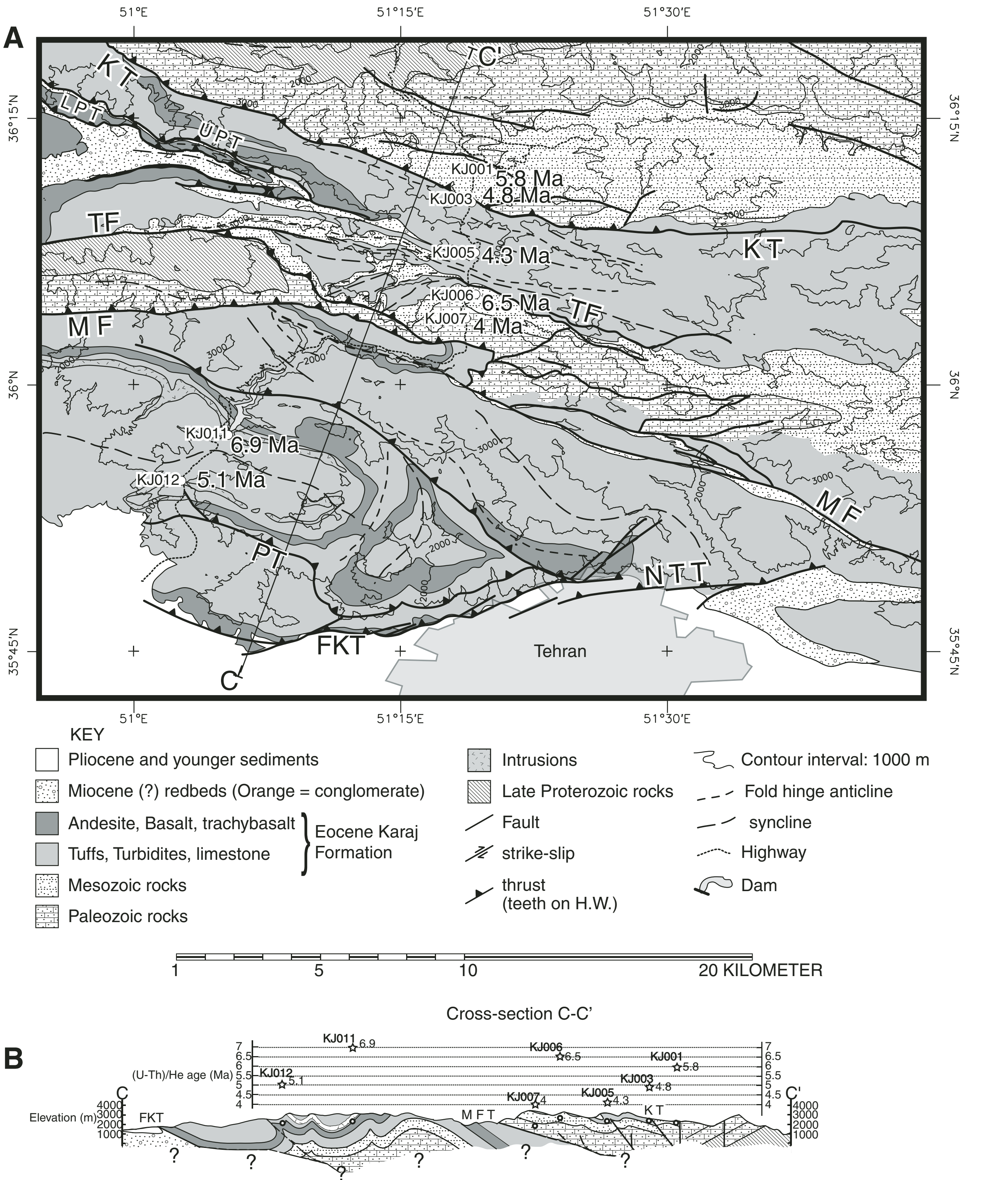


Figure 7. (A) Geologic map of the Chalus Road area, modified from the Geological Survey of Iran's Amol and Tehran 1:250,000-scale sheets (Haghipour et al., 1987; Vahdati Daneshmand, 1991). The map shows the location of samples collected along the highway, with their respective apatite (U-Th)-He cooling ages. KT—Kandavan thrust; UPT—upper Parachan thrust; LPT—lower Parachan thrust; TF—Taleghan fault; MF—Mosha fault; NTT—north Tehran thrust; FKT—Fahrahzad-Karaj thrust; PT—Purkan thrust. (B) Cross-section C–C′, with projected sample locations and (U-Th)-He cooling ages plotted above. Because we projected samples horizontally and perpendicular to the cross section, they do not necessarily plot in the sampled unit. These data show that the entire south side of the west-central Alborz was tectonically active prior to ca. 7 Ma.

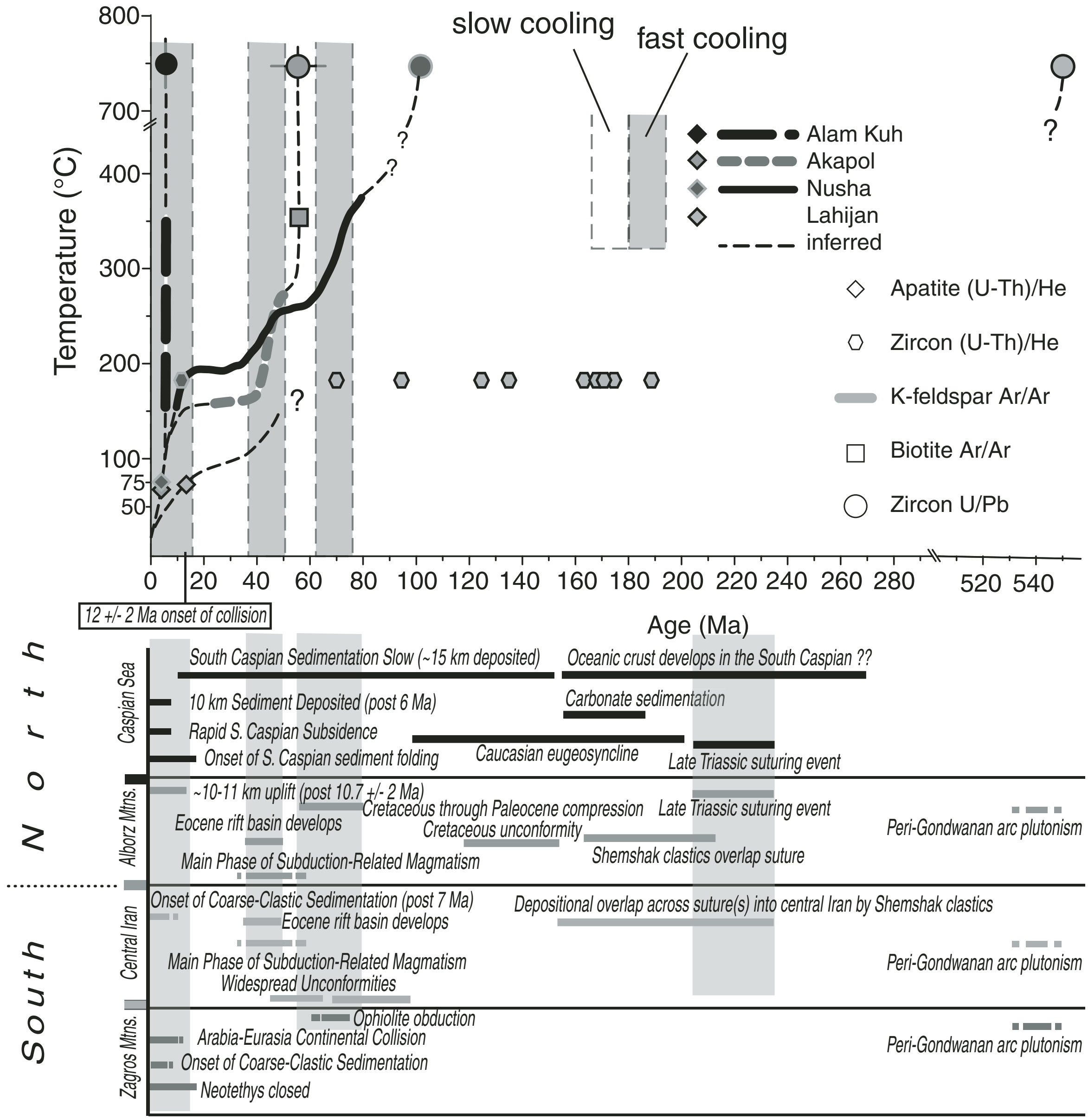


Figure 8. Summary of geochronological and thermochronological data presented in this article and in Axen et al. (2001). We show the thermal histories relative to major events or conditions known to have occurred in the Zagros, Central Iran, the Alborz, and the south Caspian Basin; references follow. The light gray areas show the temporal distribution of tectonic events in Iran (lower plot) and indicate times from which geologic evidence was obtained for orogenesis or taphrogenesis (e.g., regional unconformities, coarse clastics, deformation, etc.). The dark gray bands in the thermal history (upper) plot indicate episodes of higher cooling rate. Note that the geologic evidence for post-Cretaceous orogenesis-taphrogenesis throughout Iran generally correlates with the cooling episodes expressed by the data. References are as follows. Caspian Sea: Neprochnov (1968), Apol'skiy (1975), Degens and Paluska (1979), Berberian (1983), Zonenshain and Le Pichon (1986), Nadirov et al. (1997), Devlin et al. (1999). Alborz: Gansser and Huber (1962), Stocklin (1968, 1974a and b), Stocklin and Setudehnia (1977), Berberian (1983), Zonenshain and Le Pichon (1986), Şengör (1990), Alavi (1996), Axen et al. (2001). Central Iran: Gansser (1955), Stocklin (1968, 1974b), Stocklin and Setudehnia (1977), Berberian and King (1981), Berberian and Berberian (1981), Şengör (1990), Ramezani and Tucker (2003). Zagros: Stocklin (1968, 1974b), Stocklin and Setudehnia (1977), Berberian and King (1981), Berberian and Berberian (1981), Şengör (1990), Alavi (1994), Şengör and Natal'in (1996), Robertson (2000), McQuarrie et al. (2003), Allen et al. (2004), McQuarrie (2004).

periods identified in Iran (Fig. 8). The first post-Jurassic cooling event (recorded by K-feldspar from the Nusha pluton) occurred during Late Cretaceous to Paleocene time, a period of voluminous regional arc magmatism during which a compressional event occurred along the Neotethyan margin (Berberian and King, 1981; Berberian, 1983) (Fig. 3). In the Alborz, Late Cretaceous to Paleocene folding and uplift related to this compressional event led to deposition of the Fajan Formation conglomerate (Annells et al., 1975a; Annells et al., 1977). K-feldspar thermal history results from the Nusha pluton indicate that by 60 Ma the pulse of erosional(?) exhumation and cooling from this event had largely subsided.

The second cooling event correlates with the Eocene transgression in the southwestern Alborz region and is recorded by K-feldspar thermal history results from both the Akapol and Nusha plutons from ca. 50 to 37 Ma. The nummulitic limestone at the base of the Eocene Karaj Formation probably indicates early shallow conditions after regional flooding of the eroded former Cretaceous fold belt (Fig. 3). This was followed quickly, however, by rapid subsidence probably related to extension in the southwestern Alborz region (Stocklin, 1974a; Berberian and King, 1981; Berberian, 1983; Hassanzadeh et al., 2002; Allen et al., 2003a; Guest, 2004). This subsidence event accommodated deposition of the thick (>3 km) Karaj Formation, predominantly submarine (felsic) pyroclastic deposits interbedded with marine turbidite deposits, submarine slump deposits, and thick shale-dominated sequences (Dedual, 1967; Stocklin, 1974a; Allen et al., 2003a).

The Akapol and Nusha thermal-history-model calculations assume monotonic cooling and do not consider reheating by burial. Apparently thick Eocene Karaj basin deposits in the Alborz thin northward onto a paleo-highland south of the south Caspian Basin (Stocklin, 1974a; Berberian, 1983). Thick Eocene volcaniclastic and volcanic deposits are absent along the north flank of the Alborz (Annells et al., 1975b; Vahdati Daneshmand, 1991). Allen et al. (2003a) suggest that the Karaj Formation may have extended to the northern flank of the Alborz and was eroded during late Cenozoic orogenesis, thereby explaining the lack of outcrop. If this was the case, a thick Eocene sedimentary section (either the Karaj Formation or its nonvolcanic lateral equivalent) should be preserved in the south Caspian Basin. However, cross sections based on tying onshore stratigraphy to seismic-reflection-survey results (Huber and Eftekhar-nezhad, 1978a, 1978b) show no evidence of Eocene sedimentation in the south Caspian. Furthermore, the preserved Karaj Formation sections clearly show northward thinning of this formation in the southwestern Alborz (Stocklin, 1974a; Guest, 2004, 2006). Thus the geology of the Alborz and south Caspian indicates a paleodivide where Akapol and Nusha plutons are presently located and supports the assumption of monotonic cooling.

The combination of extension-related subsidence that created an Eocene basin and the presence of a subaerially exposed region of little to no deposition and probable erosion between the Karaj and Caspian Basins suggests a horst and graben geometry for the western Alborz region during Eocene time (e.g., Berberian, 1983). The youngest lava flows of the Eocene Karaj Formation are interbedded with the basal units of the Oligocene lower red bed sequence and mark the transition from a period of volcanism, active extension, and deep marine conditions to a second quiescent period when slow deposition in alternating shallow marine and fluvio-lacustrine conditions dominated (ca. 35 to ca. 14 Ma[?]) (Guest, 2004). K-feldspar thermal history results from both the Nusha and Akapol plutons record this period of isothermal conditions between ca. 35 and ca. 15 Ma (Fig. 8).

Collision-Related Exhumation

Ongoing orogenesis in the Alborz is linked to the Arabia-Eurasia collision by the onset of rapid sedimentation and subsidence in the south Caspian Basin (ca. 6 Ma; Nadirov et al., 1997), by onset of molasse deposition along the Zagros Ranges (ca. 6–7 Ma; Dewey et al., 1973; Beydoun et al., 1992), and by the development and subsequent deformation of late Miocene intramontane basins in the Alborz (Guest, 2004). These events, along with the preliminary thermochronological data from the Akapol and Alam Kuh plutons (compiled from Axen et al., 2001, in Fig. 8), were interpreted to record latest Miocene onset of significant collision-related orogenesis at ca. 6 to 7 Ma (Axen et al., 2001). Furthermore, Axen et al. (2001) link tectonic reorganization in the Middle East at 5 ± 3 Ma (as stated) to choking of the Neotethyan subduction zone by thick Arabian continental lithosphere.

In contrast, Robertson (2000) argues that collision occurred between 23 and 16 Ma, and Allen et al. (2003) suggest that tectonic reorganization reflects the migration of deformation (at 5 ± 3 Ma) out of regions of thickened crust (Zagros suture zone and Greater Caucasus) into the regions of little or no previous collision-related crustal thickening. New magnetostratigraphic data from syncontractional sediments preserved at the Zagros mountain-front flexure (Homke et al., 2004) (Fig. 1) show that deformation along one part of this flexure started at 7.6 ± 0.5 Ma, which implies that deformation was occurring north of the flexure prior to 7.6 ± 0.5 Ma.

Paleoceanographic constraints from carbon and oxygen isotope analysis of benthic foraminifers show that warm salty water known as Tethyan-Indian saline water was flowing into the Indian Ocean from the Tethys Ocean until 14 Ma, after which time circulation patterns shifted to the present thermohaline circulation system (Woodruff and Savin, 1989; Mohajjel et al., 2003). This suggests that the Neotethyan Ocean may have closed or become very small by 14 Ma, implying that the earliest stages of continental collision were in progress by this time.

Our data from the west-central Alborz suggest that the onset of exhumation in the Nusha area started shortly after 12 Ma (Figs. 4 and 8). In addition, this article shows middle Miocene apatite (U-Th)-He cooling ages for the Lahijan pluton on the Caspian coast (Fig. 5) and late Miocene apatite (U-Th)-He cooling ages for samples from across the Alborz (Chalus Road and Dizan transects; Figs. 6 and 7). The Nusha transect yields an apparent exhumation rate of 0.45 km/m.y (post–12 Ma). These results are consistent with the regional denudation rate for the west-central Alborz, which has varied from <0.1 km/m.y. prior to 12 Ma to ~5 km/m.y. from ca. 12 Ma until the present day. Thus the onset of rapid exhumation in the Alborz probably occurred earlier than was suggested by Axen et al. (2001). These authors showed good evidence for widespread cooling after ca. 5 Ma in the Alborz, but they could not constrain the onset of denudation.

Taken together, the data presented for northern Iran in this article, and for the Zagros mountain-front flexure by Homke et al. (2004), as well as the paleoceanographic data of Woodruff and Savin (1989), suggest that the Neotethys closed at or just prior to 14 Ma and that deformation related to the collision between Arabia and Eurasia was occurring throughout Iran by 12 Ma. This implies that collision-related deformation began approximately synchronously across a broad region that included the Zagros fold belt and the Alborz early in the collisional event rather than having migrated outward from the suture over 5–7 Ma.

Additional problems arise, however, when we place the onset of collision-related deformation in the Alborz at 12 Ma. Ignoring the problem of strike-slip deformation for the moment, shortening estimates, and shortening rates derived from global positioning system (GPS) calculations for the Alborz, are inconsistent with the 12 Ma age for the onset of deformation advocated here. The present shortening rate across the Alborz, based on a campaign GPS survey, is 5 ± 2 mm/yr (Vernant et al., 2004). In contrast, absolute shortening estimates based on restored cross sections across the Alborz are ~30 km west of Tehran (Allen et al., 2003a) and 36 ± 2 km east of Tehran (Guest et al., 2006), which imply constant shortening rates of 2.5–3.0 mm/yr, using 12 Ma as the onset of shortening.

The GPS-derived shortening rates can be reconciled with our estimate for the onset of collision-related deformation in the Alborz at ca. 12 Ma if we consider that crustal deformation has probably involved material moving into and out of the plane of the cross section (Guest et al., 2006). In Guest et al. (2006) a finite shortening model is presented that accounts for range-normal contraction and strike-slip deformation that affect the Alborz. These authors suggest that the integrated shortening across the Alborz may be as high as 53 ± 3 km, which is consistent with the shortening predicted by holding the GPS shortening rate of 5 ± 2 mm/yr constant for 12 Ma.

Alternatively, the inconsistency between the onset of deformation, shortening rate, and finite shortening could be explained by a shortening rate that has not been constant through time. For example, an arbitrary shortening rate of 1 mm/yr across the Alborz from ca. 12 to 5 Ma would result in ~7 km of shortening. If we add this to the 25 km of shortening that would have accumulated at a rate of 5 mm/yr, we get ~32 km of shortening over 5 m.y. across the Alborz since ca. 12 Ma. This estimate falls within the error of the shortening estimate made by Allen et al. (2003) and the range-normal shortening estimate presented by Guest et al. (2006). However, the very low initial shortening rates (1 mm/yr) required for this alternative to yield reasonable results would allow only a few kilometers of shortening to have accumulated between 12 and 7 Ma and would therefore probably not have caused rapid exhumation and cooling.

Thus we adopt aspects of the ideas put forth by both Axen et al. (2001) and Allen et al. (2004). Closure of the Neotethys probably occurred at ca. 14–12 Ma, resulting in deformation across much or all of present-day Iran from the Zagros simply folded belt to the developing Turkish-Iranian Plateau and the high western Alborz. The 5 ± 3 Ma tectonic reorganization probably started when thicker buoyant continental crust began to be subducted beneath Eurasia, thereby choking the subduction zone.

This interpretation may explain the inconsistency between the timing (ca. 5 Ma) of the onset of rapid south Caspian subsidence observed offshore of Azerbaijan and Turkmenistan (Nadirov et al., 1997) and the timing of rapid exhumation in the Alborz. Sedimentation rates in the south Caspian increased from <200 m/m.y. before 6 Ma to ~3000 m/m.y. by 1 Ma (Nadirov et al., 1997). Increased deformation and exhumation at 5 ± 3 Ma that affected the region surrounding the south Caspian Basin (Alborz, Greater and Lesser Caucasus, and Kopet Dagh Mountains) probably resulted in the observed increase in sedimentation rates. It is difficult, however, to explain the lack of an increase in sedimentation rates at 12 Ma, when we propose that deformation and exhumation in the western Alborz began. The sedimentation rate estimates of Nadirov et al. (1997) show a gradual increase from ~100 m/m.y. to ~200 m/m.y. between 100 and 10 Ma, and then rates remained constant until ca. 6 Ma. Several possible explanations account for this apparent inconsistency.

The first is that sedimentation rate estimates for the south Caspian come from the northern half of the south Caspian (Azerbaijani and Turkmenistani waters) and may not accurately reflect the sedimentation rate history for the part of the south Caspian Basin that lies adjacent to the Alborz. It is therefore possible that this portion of the south Caspian Basin contains a record of high sedimentation rates for the middle Miocene.

Alternatively, paleoclimate may be important. If the Nadirov et al. (1997) estimates apply to the entire south Caspian Basin, then perhaps the climate at 12 Ma was arid, causing low erosion rates such that the sediment supply did not reflect the tectonic activity occurring at the time. Sedimentologic and stratigraphic evidence shows that middle to late Miocene time was a period of evaporite deposition throughout southern, central, and northern Iran (Stocklin and Setudehnia, 1977), suggesting a hot, arid climate.

A third explanation for the lack of an increase in sedimentation rate in the south Caspian at 12 Ma could be that much eroded material did not reach the south Caspian. Perhaps early drainage basins that fed the south Caspian were small. As deformation progressed, the drainage basins in the Alborz probably grew and captured axial drainage systems within intramontane basins in the deformation belt. The Sefid River, which presently cuts across the western Alborz, probably began to drain central Iran south of the Alborz by latest Miocene to Pliocene–Pleistocene time (Annells et al., 1975a). It is clear that during Miocene time, large volumes of sediment were stored in intramontane basins in the southern Alborz (e.g., Gand Ab and Narijan red bed sedimentary units described by Guest, 2004) and that the bulk of these deposits was deformed and erosionally removed since late Miocene time (Guest, 2004). Headward erosion of north Alborz drainage systems into large volumes of poorly consolidated sediment in the southern Alborz and central Iran may have contributed to a sediment pulse in the south Caspian long after deformation began.

Finally, the temporal evolution of shortening rates in Iran remains unconstrained. It is possible, therefore, that deformation in northern Iran was initially occurring at a low enough rate as not to affect the northern south Caspian and that the sediment pulse observed in the south Caspian at 5 Ma reflects an increase in deformation rate just prior to 5 Ma.

The inconsistencies described previously are partly due to incomplete data from the Alborz and Iran in general. Additional thermochronometric, geochronologic, structural, and sedimentological studies in Iran, Turkey, Afghanistan, Azerbaijan, and the south Caspian are required to understand the collisional processes acting in these regions.

CONCLUSIONS

Thermochronological and geochronological data presented in this paper are consistent with three significant tectonic events and two quiescent periods identified in the post-Jurassic geologic record of the Middle East. Our thermochronological data resolve cooling events separated by isothermal periods that correlate with (1) Cretaceous to Paleocene orogenesis related to compression along the Neotethyan margin, (2) latest Paleocene to earliest Eocene shallow marine carbonate and evaporite sedimentation over the rocks deformed and eroded in the previous orogenic event, (3) Eocene intra-arc–back-arc extension in the Alborz region, (4) transgression of the Oligocene to earliest middle Miocene Qom Sea over central Iran and the southern Alborz, and (5) the middle Miocene collision between the Arabian subcontinent and the Turkish-Iranian active continental margin. Furthermore, these data provide the best available estimate for the onset of late Cenozoic orogenesis in the western Alborz (ca. 12 Ma) and suggest that deformation was occurring approximately synchronously across the present width of the western Alborz by latest Miocene time and that faults along the southern range margin have therefore been active since at least this time.

Finally, we suggest that middle Miocene sedimentation along the southern margin of the Zagros simply folded belt (Homke et al., 2004), the disappearance of the Neotethys Ocean (Woodruff and Savin, 1989), the onset of rapid cooling and exhumation in the Alborz (this article), and Turkish-Iranian Plateau uplift inferred from sedimentation patterns (Dewey and Şengör, 1979; Şengör and Kidd, 1979) are linked to the Arabia-Eurasia collision. All of this suggests that continent-continent collision zones like the Indo-Asia and Arabia-Eurasia collisional belts can develop as broad mobile regions of distributed deformation sandwiched between strong immobile converging cratons. Continued post-collision convergence is accommodated by lateral escape of lithospheric fragments (e.g., the Anatolian microplate escaping to the west), by lithospheric thickening (e.g., lithospheric thickening occurring beneath the high Zagros and Tibet), or by crustal flexure (e.g., northern Iran and the south Caspian; Guest, 2004).

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