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Timing of uplift in the Zagros belt/Iranian plateau and accommodation of late Cenozoic Arabia–Eurasia convergence

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Abstract – The motion of Arabia was stable with respect to Eurasia over the past 22 Ma. Deformation and exhumation in the Zagros is seen to initiate at the same time as argued by new detrital thermochronologic constraints and increasing accumulation rates in synorogenic sediments. A recent magnetostratigraphic dating of the Bakhtyari conglomerates in the northern Fars region of the Zagros further suggests that shortening and uplift in the Zagros Folded Belt accelerated after 12.4 Ma. Available temporal constraints from surrounding collision belts indicate that shortening and uplift focused in regions bordering the Iranian plateau to the south between 15 and 5 Ma. As boundary velocity was kept constant this requires concomitant decreasing strain rates in the Iranian plateau. Slab detachment has been proposed to explain the observed changes as well as mantle delamination, but the insignificant change in the Arabian slab motion and lack of unambiguous constraints make both hypotheses difficult to account for. It is proposed based on a review of shortening estimates provided throughout the Arabia–Eurasia collision that the total 440 km of convergence predicted by geodesy and plate reconstruction over the past 22 Ma can be accounted for by distributed shortening. I suggest that the topography and expansion of the Iranian plateau over Late Miocene–Pliocene time can be reproduced by the progressive thickening of the originally thin Iranian continental lithosphere presumably thermally weakened during the Eocene extensional and magmatic event.

Keywords: Zagros, plateau, uplift, shortening.

1. Introduction

Knowledge of the distribution of Cenozoic shortening in the Zagros collision in Iran is critical to better understand how the Arabian plate motion was accommodated during the collision with the overriding Eurasian plate. Combined with the precise timing of deformational events, it is key in linking the kinematic development of the Zagros Folded Belt to the growth of the Iranian plateau.

A significant number of publications have brought new insights on the current and Quaternary tectonics of the Zagros mountain belt (Nilforoushan *et al*. 2003; Masson *et al*. 2005; Vernant *et al*. 2004; Walpersdorf *et al*. 2006; Oveisi *et al*. 2009) and on the deep geophysical settings beneath the Iranian plateau and the Zagros belt (Hatzfeld *et al*. 2003; Maggi & Priestley, 2005; Paul *et al*. 2006; Kaviani *et al*. 2009; Hatzfeld & Molnar, 2010). Allen, Jackson & Walker (2004) pointed out that major reorganization of the Arabia–Eurasia collision has occurred in the past 5 ± 2 Ma to account for the rates of motion along major active faults. However, thermochronometric data (Fig. 1) from the Zagros foreland sediments argued for acceleration of denudation *c*. 25 Ma (Homke *et al*. 2010; S. Khadivi, unpub. Ph.D. thesis, Univ. Pierre et Marie Curie, 2010), and thrusting/folding activity in the northern Zagros belt seems to have been mostly initiated in Early–Middle Miocene time (Gavillot *et al*. 2010; Khadivi *et al*. 2010). Overall, constraints from the Zagros are rather in agreement with the stable northward drift of the Arabian plate since 22 Ma (ArRajehi *et al*. 2010).

The deep structure (Fig. 2) shows a 45 km thick Arabian crust beneath the Zagros Folded Belt and the High Zagros (Paul *et al*. 2006, 2010). The good agreement with the unthickened portion of the Arabian margin (Gök *et al*. 2008) indicates that the crust has not yet been significantly thickened beneath the Zagros Folded Belt. By contrast, the deepening of the Moho to a depth of 70 km beneath the Sanandaj–Sirjan Metamorphic Belt illustrates the significant underthrusting of the Arabian margin and the focused accretion by underplating beneath the upper Iranian plate (Fig. 2). The thickening of the lithosphere is supported by seismological evidence indicating that there is a thick lithosphere 'core' beneath the Zagros (Priestley & McKenzie, 2006). North of the Sanandaj–Sirjan Zone, the Iranian continental block displays a crustal thickness of ∼ 50 km and a warm upper mantle lithosphere down to a depth of 100 km. This anomalously thin lithosphere might be caused by the partial delamination of a continental lithosphere following the thickening of the continent during the protracted plate convergence (Maggi & Priestley, 2005; Hatzfeld & Molnar, 2010). But a more accurate

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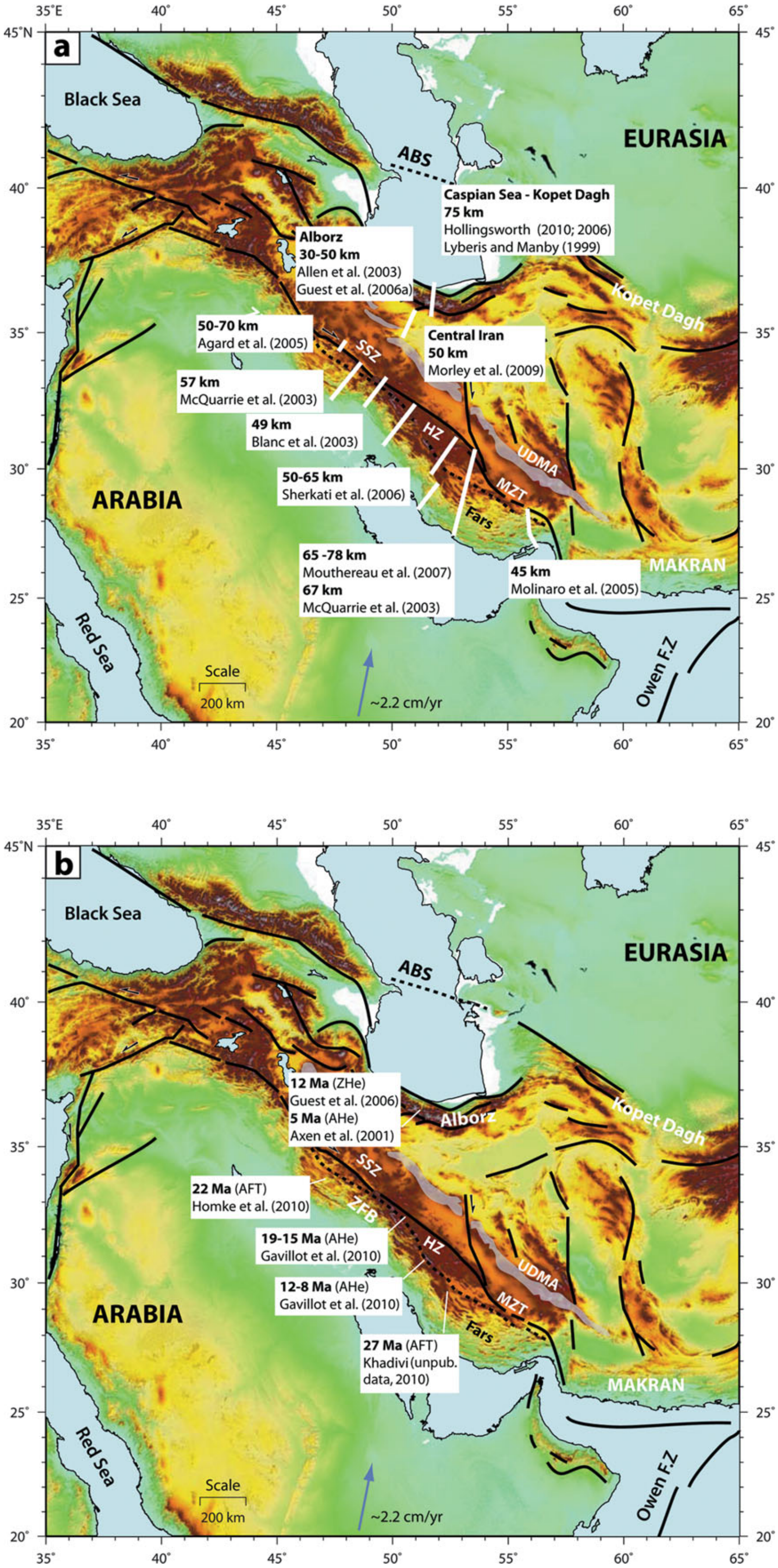

Figure 1. (Colour online) (a) Geodynamic setting of the Arabia–Eurasia collision and the distribution of long-term shortening and (b) ages of the most recent exhumational events according to the thermochronometer used (AFT – apatite fission-track; AHe – (U–Th)/He dating on apatite; ZHe – (U–Th)/He dating on zircon). Main topographic and tectonic features of the Arabia–Eurasia convergence are also shown. White lines correspond to the location of balanced cross-sections from which amounts of shortening have been estimated. Black lines display major active faults. The current Arabian–Eurasian plate convergence is shown as a grey (blue) arrow after Vernant *et al*. (2004). Abbreviations are Zagros Folded Belt (ZFB), High Zagros (HZ), Main Zagros Thrust (MZT), Sanandaj–Sirjan Zone (SSZ), Urumieh–Dokhtar Magmatic Arc (UDMA), Apsheron–Balkan Sill (ABS).

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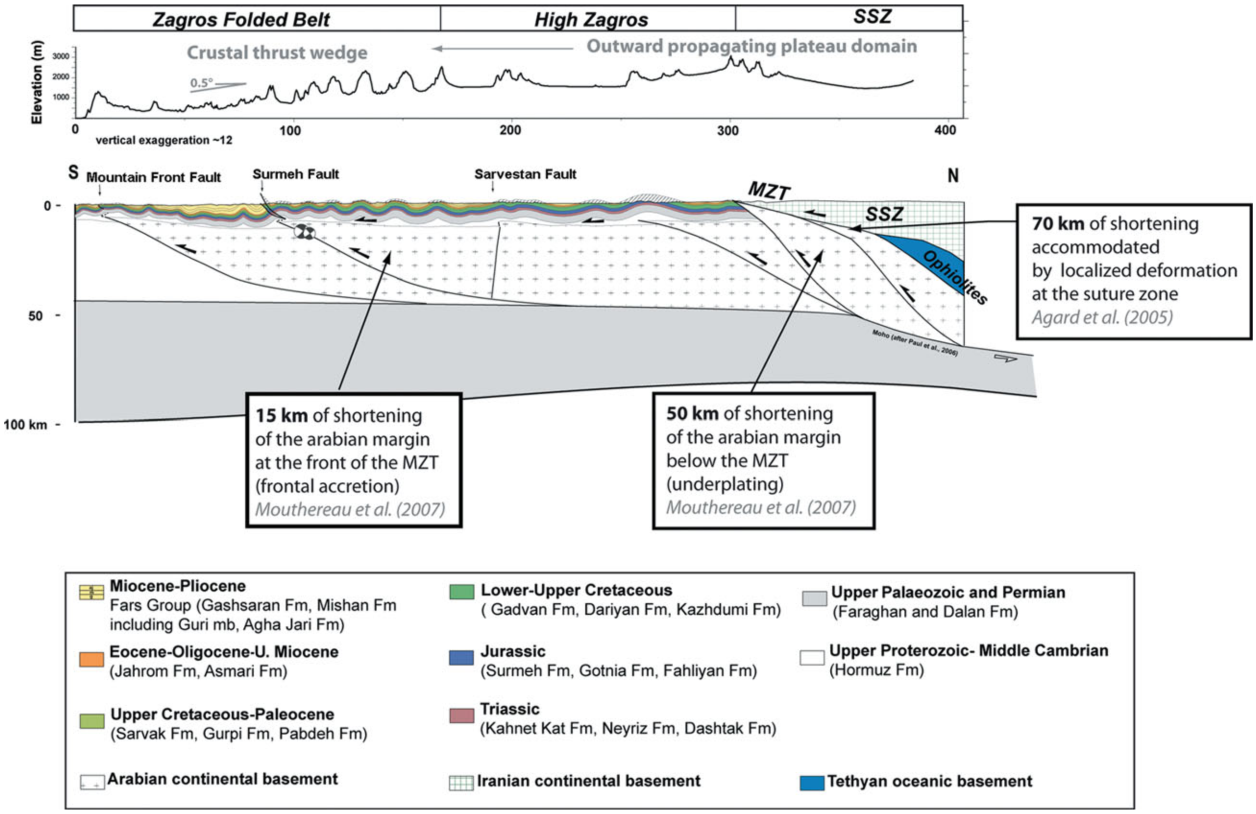

Figure 2. (Colour online) Distribution of shortening across the Zagros belt and outward migration of plateau uplift. The balanced cross-section of the Zagros in the Fars region is after Mouthereau *et al*. (2007). See Figure 1 for location and abbreviations.

velocity estimate does not support mantle delamination (Kaviani *et al*. 2007), and more generally there is no definitive evidence supporting the convective removal of lithosphere beneath the plateau. On the other hand, upwelling of asthenospheric mantle controlled by slab retreat may provide an explanation for such thin lithosphere as suggested by several geological constraints (Vincent *et al*. 2005; Verdel *et al*. 2007; Morley *et al*. 2009).

A kinematic link between the recent tectonic evolution of the Zagros Folded Belt and the Iranian plateau growth can be suggested based on several lines of evidence. The southern edge of the Iranian plateau is coincident, in the Fars region of Iran, with the northern edge of the Zagros Mountains outlined by a cumulative topographic step and structural elevation of ∼ 2 km (Figs 2, 3). Such a morphology indicates that the regional Zagros topography was built by basement thrust units, the most active ones being spaced ∼ 80 km apart (Mouthereau, Lacombe & Meyer, 2006; Mouthereau *et al*. 2007). Combined with evidence of widespread seismicity over the length of the outer Zagros Folded Belt, this supports a model in which the topography is balanced by a crustal-scale critically tapering orogenic wedge.

By contrast, the High Zagros region forms an elevated low-relief area that is morphologically not distinguishable from the southern edge of the Iranian plateau (Figs 2, 3). This suggests that part of the Zagros collision belt has been uplifted owing to its incorporation into the Iranian plateau. This relationship implies that the growth history of the plateau cannot be understood outside the context of the kinematic history of the Zagros Folded Belt.

In this short paper, by providing a review of the recent advances on the temporal evolution and spatial distribution of shortening and exhumation in the Zagros belt and other compressional domains surrounding the Arabia–Eurasia collision, I aim at highlighting the timing and mechanisms of Iranian plateau growth. Specifically, I focus on the distribution of shortening over the past 22 Ma, a period during which the northward motion of Arabia was stable.

2. Regional geological background

The NW–SE-trending Zagros orogeny, which is part of the much larger Alpine–Himalayan orogenic system, extends some 2000 km from the East Anatolian fault in eastern Turkey to the Makran subduction in southern Iran (Fig. 1). A GPS-derived velocity model shows present-day convergence rates between Arabia and Eurasia of 19–26 mm yr^{-1} (McClusky *et al*. 2003; Vernant *et al*. 2004). In the next Sections, I briefly present the main geological features of the Zagros collision including the Zagros belt, the Sanandaj–Sirjan belt and the Urumieh–Dokhtar volcanic arc.

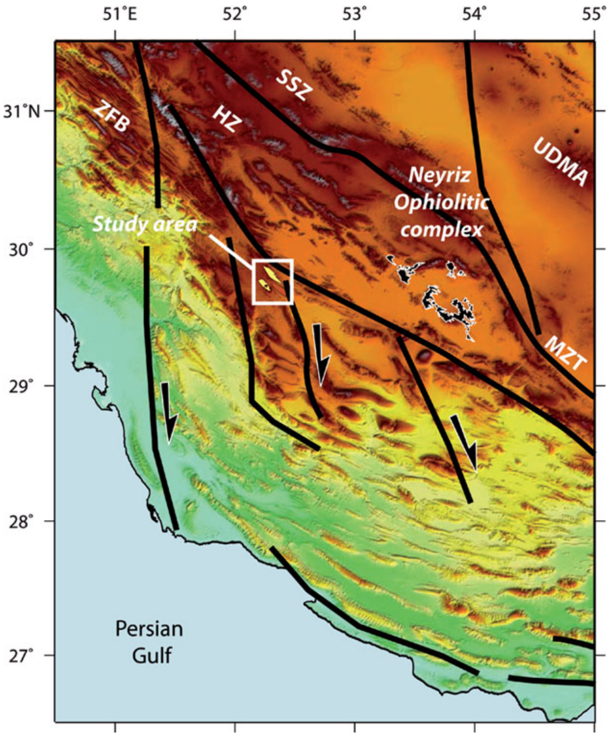

Figure 3. (Colour online) Topographic map of the Fars area (SRTM 90 m digital elevation data; http://srtm.csi.cgiar.org) showing the location of the area studied for magnetostratigraphy and thermochronometry (Derak anticline) by Khadivi *et al.* (2010) and S. Khadivi (unpub. Ph.D. thesis, Univ. Pierre et Marie Curie, 2010). The Neyriz Ophiolitic Complex is currently exposed as klippen above the deformed sedimentary units of the High Zagros (HZ). The metamorphic belt of the Sanandaj–Sirjan Zone (SSZ), the Urumieh–Dokhtar Magmatic Arc (UDMA), the Zagros Folded Belt (ZFB) and Main Zagros Thrust (MZT) are also labelled.

2.a. Zagros Folded Belt (ZFB)

The Zagros Folded Belt makes up the currently active accretionary wedge of the Zagros collision. It is characterized by remarkably regular, long and large-wavelength NW-trending concentric folds (Figs 2, 3). They have probably resulted from buckling and subsequent detachment folding of a 12 km thick sediment cover enabled by the detachment in the Cambrian Hormuz salt (Lacombe *et al.* 2007; Mouthereau *et al.* 2007). Active faulting is rare but does occur in the competent cover as argued from recent seismological studies (Adams *et al.* 2009; Nissen *et al.* 2010; Roustaei *et al.* 2010). The pre-Cambrian basement of the Arabian margin is also actively deforming, as indicated by a number of morphotectonic observations in the Fars (Molinaro *et al.* 2004; Lacombe *et al.* 2006; Mouthereau *et al.* 2007) and seismicity (Talebian & Jackson, 2004). Basement-involved shortening is also mechanically required to maintain the regional topography (e.g. Mouthereau, Lacombe & Meyer, 2006) and it is confirmed by the most recent analysis of individual earthquakes revealing active reverse faulting at depths of 10–30 km (Roustaei *et al.* 2010).

The external Zagros can be divided in two sub-structural domains. The first one is the High Zagros (HZ) belt characterized, in the Fars region, by Mesozoic carbonates overthrust by the radiolaritic series and ultramafic bodies of the Neyriz ophiolitic complex, considered allochthonous fragments of the western Neo-Tethyan ocean (Figs 2, 3) (Stocklin, 1968; Golonka, 2004). The second is the Zagros Folded Belt (ZFB) *sensu stricto*, also called the Zagros Simply Folded Belt (ZSFB), with folded Miocene to Pliocene synorogenic strata (Fig. 2).

2.b. Sanandaj–Sirjan Zone (SSZ)

The Sanandaj–Sirjan Zone, located to the north of the Main Zagros Thrust (MZT), represents the internal tectonomagmatic and metamorphic part of the Zagros belt (Figs 1–3).It is made of sedimentary and metamorphic (HP/LT and HT/LP facies) Palaeozoic to Cretaceous rocks formed in an accretionary prism located to the south of the Iranian microcontinent separated from Gondwanaland during Late Jurassic time (Berberian & Berberian, 1981; Golonka, 2004). Alternative interpretations consider it to be the metamorphic core of a larger Zagros accretionary complex built by the thickening of distal crustal portions of the Arabian margin (Shafaii Moghadam, Stern & Rahgoshay, 2010). During the second half of the Mesozoic (Middle Jurassic–Early Cretaceous), part of the Sanandaj–Sirjan Zone was an active Andean-like margin characterized by calc-alkaline magmatic activity in which mainly andesitic and gabbroic intrusions were emplaced (Berberian & Berberian, 1981). Magmatism resumed in Paleocene–Eocene time, as evidenced by gabbroic intrusions (Leterrier, 1985; Mazhari *et al.* 2009) or granitic intrusions of this age (Rachidnejad-Omran *et al.* 2002).

2.c. Urumieh–Dokhtar Magmatic Arc (UDMA)

The Urumieh–Dokhtar Magmatic Arc (UDMA; Fig. 1) is interpreted as a subduction-related arc that has been active from Late Jurassic time to the present (Berberian & King, 1981; Berberian *et al.* 1982). The climax of magmatic activity can be dated to Middle Eocene time (Berberian & King, 1981). The volcanic rocks of the Urumieh–Dokhtar Magmatic Arc are composed of voluminous tholeiitic, calc-alkaline and K-rich alkaline magmatic rocks with associated pyroclastic and volcanoclastic successions. Magmatism resumed in Pliocene time and the Quaternary as indicated by lavas and pyroclastic rocks associated with the volcanic cones of alkaline and calc-alkaline nature (Berberian & Berberian, 1981). The Plio-Quaternary volcanism was suggested to result from the modification of geothermal gradients that was tentatively related to lithosphere delamination beneath the Iranian plateau (Hatzfeld & Molnar, 2010) or slab break-off (Omrani *et al.* 2008).

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3. Timing of shortening, collision and uplift in the Zagros belt

3.a. Short-term, long-term shortening and the Arabia–Eurasia convergence

Comparison between a recent synthesis of GPS data (ArRajehi *et al*. 2010) and reconstruction of past plate motions (McQuarrie *et al*. 2003) shows that the Arabia–Eurasia convergence occurred at a rate of ~ 20 km Ma^{-1} (Tatar *et al*. 2002; Hatzfeld *et al*. 2003; Nilforoushan *et al*. 2003; Vernant *et al*. 2004) since at least 22 Ma, following the separation of Arabia from Africa (Nubia), the onset of rifting in the Red Sea and the Aden Gulf and the increase in plate coupling in the Zagros collision (e.g. Mouthereau *et al*. 2007).

A total convergence of 440 km should have been accommodated by distributed collisional shortening and subduction (i.e. underthrusting of the continental lithosphere) in the surrounding collision belts since 22 Ma including the Zagros to the south, the Alborz and the Kopet-Dagh to the north, and by N–S shortening accommodated by reverse and/or strike-slip faulting in Central Iran (e.g. Allen *et al*. 2011 and references therein).

For the Zagros alone, geodetic measurements argue for current shortening rates of 7–10 mm yr^{-1} (Tatar *et al*. 2002; Nilforoushan *et al*. 2003; Vernant *et al*. 2004), with most of the current shortening accumulating within the lower elevation parts of the Zagros Folded Belt (Walpersdorf *et al*. 2006) in agreement with geomorphological observations (Oveisi *et al*. 2009), thus fitting the seismicity distribution well. By comparison, all published balanced cross-sections, irrespective of differences in structural interpretations (Blanc *et al*. 2003; McQuarrie, 2004; Sherkati & Letouzey, 2004; Molinaro *et al*. 2005; Mouthereau *et al*. 2007), account for as much as 50–70 km of shortening. By assuming that the initiation of shortening dates back to 22 Ma, such a shortening accounts for less than half the current shortening rates. On the other hand, a finite shortening of 70 km would be achieved in ~ 7 Ma to be consistent with the current shortening rates. Based on these geodetic data, Allen, Jackson & Walker (2004) therefore inferred that the main episode of crustal thickening in the Zagros should be more recent than 7 Ma. However, because of the stability of the Arabian plate motion since 22 Ma (McQuarrie *et al*. 2003; ArRajehi *et al*. 2010), forces related to the assumed changes at ~ 5 Ma must have been limited because they did not alter the slab pull forces acting on the Arabian plate motion. In this context, the timing of development of the High Zagros hence appears key in constraining the Late Cenozoic distribution of shortening in the Arabian–Eurasian plate convergence and the mechanism of Iranian plateau growth. In the next Sections, I specifically explore constraints on the collision onset, the timing of deformation in the Zagros belt and the temporal evolution of exhumation in the High Zagros.

3.b. Initiation of Arabia–Eurasia collision

The Arabian and Eurasian plates started to collide along the Bitlis thrust zone in Early Miocene time (*c*. 20 Ma) following the consumption of the last remaining oceanic lithosphere (Okay, Zattin & Cavazza, 2010). Along the Zagros suture zone, the stratigraphic/structural relationships also argue for final closure of the Neo-Tethyan ocean by Early Miocene time *c*. 20 Ma (e.g. Agard *et al*. 2005). This is in line with evidence supporting the coeval onset of foreland subsidence (Mouthereau *et al*. 2007) and stress build-up in the Arabian platform (Ahmadhadi, Lacombe & Daniel, 2007). Consistently, the recent re-evaluation of the stratigraphy of the coarse-grained facies in the Zagros foreland basin shows that the onset of coarsening-upward sedimentation linked to the exhumation of the hangingwall of the Main Zagros Thrust occurred during Late Oligocene–Early Miocene time (Fakhari *et al*. 2008). This is also indicated by the finding of Mesozoic to Eocene detrital apatite fission-track (AFT) cooling ages in Miocene foreland sediments compatible with the Sanandaj–Sirjan Zone cooling history (S. Khadivi, unpub. Ph.D. thesis, Univ. Pierre et Marie Curie, 2010; see also Fig. 6). On the other hand, the decrease in or end of magmatism in Central Iran supports that initial collision of Arabia occurred in Late Eocene time (e.g. Vincent *et al*. 2005; Allen & Armstrong, 2008). On the Arabian margin, a Middle Eocene–Late Oligocene or Late Eocene–Early Miocene unconformity recognized in the carbonaceous sediment succession of the Zagros (James & Wynd, 1965; Berberian & King, 1981) and the erosional or non-depositional hiatus described to the NW, in the Lorestan area, in the Middle–Late Eocene interval (Homke *et al*. 2009) indirectly support this timing. In summary, constraints on the timing of Neo-Tethyan ocean consumption, Zagros sediment provenance and arc magmatism in the Iranian microplate support initiation of the Arabia–Eurasia collision between 35 and 20 Ma.

3.c. Timing of deformation in the Zagros Folded Belt

The unambiguous dating of deformation in the fold–thrust belt requires the preservation of tectonic/stratigraphic relationships such as synfolding sediments and associated geometries like growth strata. This is only possible in regions where regional subsidence and sedimentation supplied by exhuming mountain ranges are high enough to allow wedge-top basins to develop. Such geometries are observed in some parts of the Zagros and when combined with magnetostratigraphy allow accurate determination of the age of deformation as presented in recent papers (Homke *et al*. 2004; Khadivi *et al*. 2010).

Hereafter, I focus on the dating of the first synorogenic deposits in the northern Zagros. The studied sections are located (Fig. 4) on the northern flank of the Chahar–Makan syncline at an altitude of ~ 2500 m,

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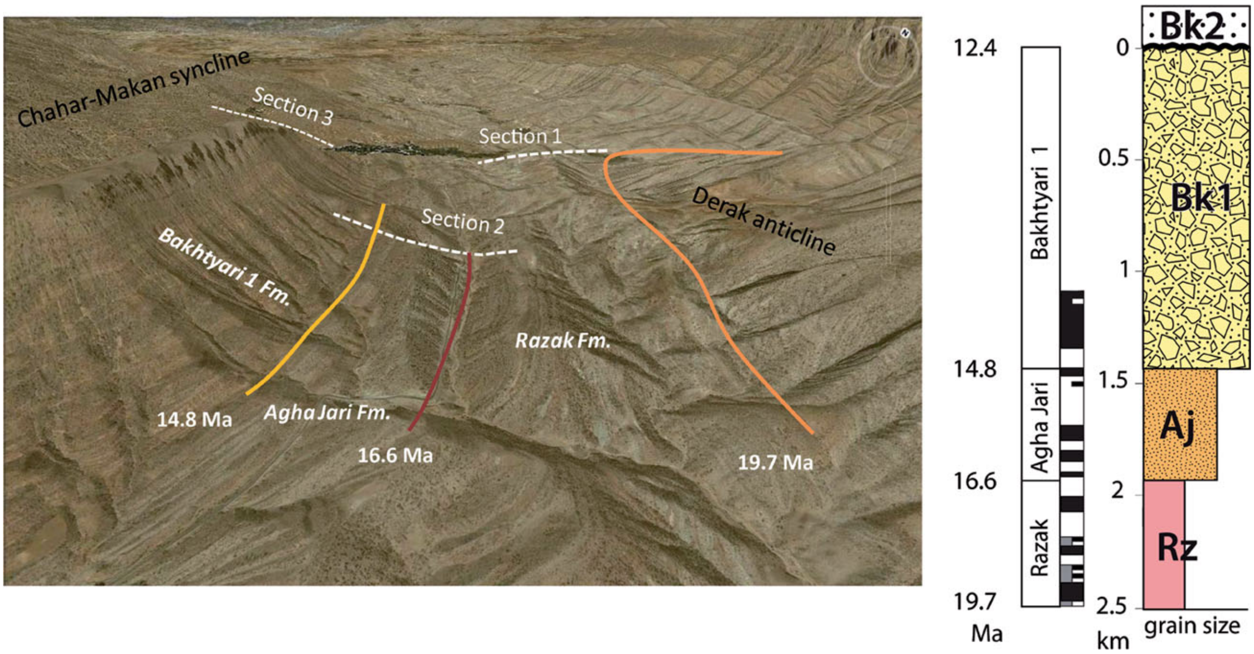

Figure 4. (Colour online) Position of magnetostratigraphic sections measured in the northern flank of the Chahar–Makan syncline and age of the main formation boundaries obtained after Khadivi *et al*. (2010). On the left, sections are shown on 3D satellite view of the studied area (See Fig. 3 for location). On the right, the total sedimentary section 2.5 km thick is shown with age constraints. The age of the youngest Bakhtyari 1 conglomerate is derived from the accumulation rates obtained from magnetostratigraphy (modified after Khadivi *et al*. 2010).

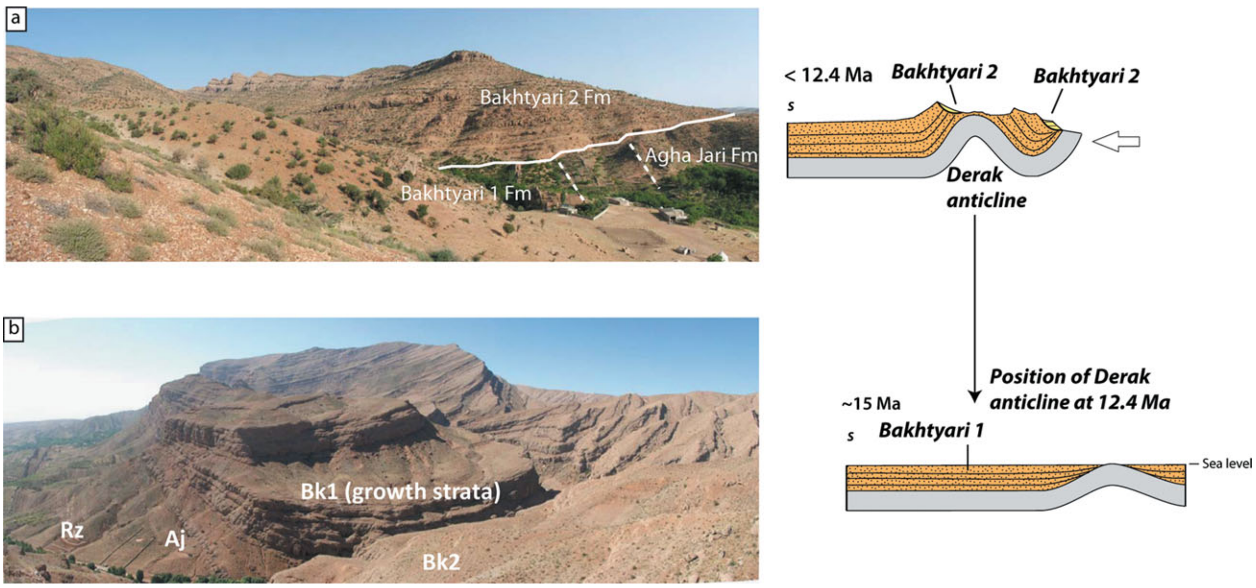

Figure 5. (Colour online) (a) Structural relationships between Bakhtyari 2 (Bk2) and Bakhtyari 1 (Bk1) conglomerates and (b) growth strata geometry on the northern flank of the Derak anticline. Interpretation of these geometries in terms of the sequence of folding is given on the right-hand side.

20 km to the NW of Shiraz, in the Fars province of Iran. The lowest strata, 500 m thick, are sediments deposited in a coastal sabkha environment and correspond to the Razak Formation, the base of which is dated to 19.7 Ma. Above are the 400 m thick deltaic sandstones of the Agha Jari Formation dated to 16.6 Ma in agreement with the finding of the nannoplankton association that indicates the NN4 biozone. Above, the lowest Bakhtyari 1 unit is made of alluvial conglomerates deposited close to sea-level, as revealed by the underlying marine Agha Jari sediments and by marine incursions in the Oligocene–Miocene Bakhtyari conglomerates deposited in the High Zagros (Fakhari *et al*. 2008; Gavillot *et al*. 2010). Growth strata found on the northern flank of the Derak anticline confirms that the Bakhtyari conglomerates were deposited during folding, therefore providing a minimum age of 14.8 Ma for the onset of folding in the northern Zagros belt

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(Fig. 5). However, this stage of deformation does not represent the main stage of folding as the Razak Fm, Agha Jari Fm and the Bakhtyari 1 Fm have been tilted by the subsequent growth of the Derak fold and are currently cropping out in the Chahar–Makan and Qalat synclines. This second folding is outlined by a major angular unconformity between the flat-lying or slightly N-dipping conglomeratic layers of the Bakhtyari 2 Formation and underlying Bakhtyari 1 Formation. By considering the total cropping-out thickness of Bakhtyari 1 conglomerates and extrapolating with accumulation rates derived from magnetostratigraphy, I obtained a maximum age of 12.4 Ma for the second major stage of folding. Taking into account age uncertainties on the unconformity, this age appears not significantly different from other magnetostratigraphic constraints obtained for folding initiation at the mountain front dated at 7.6 Ma in the Lorestan area (Homke *et al*. 2004) or from the inner Zagros belt where folding is dated to 11 Ma (H. Emami, unpub. Ph.D. thesis, Univ. de Barcelona, 2008). In the hangingwall of the Dinar thrust (High Zagros), detrital apatite (U–Th)/He ages of 11.6–8.8 Ma on folded Bakhtyari conglomerates (Gavillot *et al*. 2010) provide indirect constraints on the age of deformation. Overall, stratigraphic constraints reveal that shortening was initially accumulated in the northern Zagros in Early Miocene time, close to the suture zone, and subsequently propagated southward during latest Miocene time.

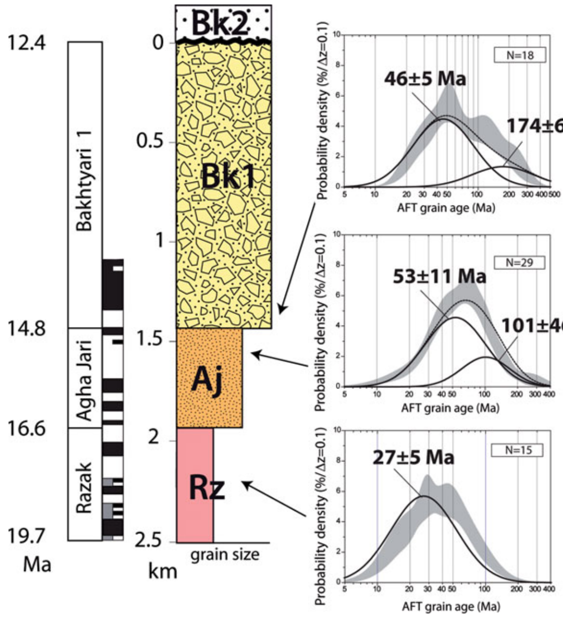

Figure 6. (Colour online) Probability density distribution of fission-track ages obtained on detrital apatites (N is the number of grains) from the Miocene sediments of the Chahar–Makan section presented in Figure 4 (modified after S. Khadivi, unpub. Ph.D. thesis, Univ. Pierre et Marie Curie, 2010) and dated by Khadivi *et al*. (2010). All grain-age populations are interpreted as cooling ages and as such indicate exhumational events. The age at 27 Ma is interpreted to be related to the rapid exhumation owing to thickening associated with the Zagros collision. Eocene and Mesozoic ages correspond to grains cooled in the Sanandaj–Sirjan Metamorphic Belt and deposited into the Miocene foreland basin, thus revealing the suturing along the Main Zagros Thrust and the onset of the Zagros collision.

3.d. Uplift and exhumation in the Zagros Folded Belt and the High Zagros

In addition to dating deformation in the Zagros, it is equally important to track the elevation changes back in time. Based on the youngest marine sediments dated in Iran, it is beyond doubt that both the Zagros and the Iranian plateau were still below sea-level until Early Miocene time (Schuster & Wielandt, 1999; Harzhauser *et al*. 2007), and one can also be confident that until $\sim$ 15 Ma the northern Zagros Folded Belt was close to sea-level (Khadivi *et al*. 2010).

Helium dating on detrital apatites from the Bakhtyari conglomerates deposited in the High Zagros and an age-elevation profile of the Lajin thrust (Fig. 1b) tells us that rapid cooling took place in Early Miocene time from 19 Ma to 15 Ma (Gavillot *et al*. 2010). Furthermore, the pre-collisional zircon (U–Th)/He ages presented in the same study indicate that the maximum exhumation in the High Zagros was limited to 7–9 km, which is consistent with the average thickness of the Meso-Cenozoic sediment cover and the scarcity of Palaeozoic rocks cropping out in the High Zagros. They deduced from the hangingwall of High Zagros thrusts local exhumation rates of the order of 0.3–0.4 km Ma^{-1}.

Low-temperature AFT thermochronology carried out on older Miocene foreland sediments of the Zagros Folded Belt (Figs 1b, 6) indicates that rapid cooling occurred between 27 Ma (depositional age of the Razak Fm is 19.7 Ma in the Chahar–Makan syncline) and 22 Ma (depositional age of the Lower Agha Jari Fm is 12.8 Ma in the Zarrinabad syncline) in the High Zagros (Homke *et al*. 2010; S. Khadivi, unpub. Ph.D. thesis, Univ. Pierre et Marie Curie, 2010). Taking into account a closure temperature of 110 °C and a geotherm of 15–24 °C km^{-1} (Mouthereau, Lacombe & Meyer, 2006; Gavillot *et al*. 2010; Homke *et al*. 2010), one estimates that 4.5–7 km were exhumed during Early Miocene time.

The preservation of unreset Mesozoic, Eocene or Early Miocene grain-age populations limits the exhumation in the Chahar–Makan syncline to 2.5 km, which is the thickness of the synorogenic Miocene sediments (Fig. 6). Since folding started later than 12.4 Ma, one can derive a minimum exhumation rate of 0.2 km Ma^{-1}, comparable to the sedimentary accumulation rates of $\sim$ 0.2–0.3 km Ma^{-1} in the 12–3 Ma distal foreland basin succession at the mountain front (Homke *et al*. 2004) and rates of 0.2–0.6 km Ma^{-1} in the 20–14 Ma proximal foreland sediments (Khadivi *et al*. 2010). Taking into consideration the fact that accumulation rates are underestimated because decompaction is not accounted for, I see no significant

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difference between erosion and sedimentation rates during the Miocene.

To summarize, thermochronologic data from Miocene sediments show rapid exhumation near the suture zone after 25 Ma (Figs 1b, 6). As a consequence this region was actively uplifting above sea-level owing to the thickening of the Arabian crust. Further evidence of exhumation at this time in the Sanandaj–Sirjan Zone is provided by the occurrence of detrital zircons derived from the overriding Iranian microplate and deposited in the Upper Oligocene conglomerates (Horton *et al.* 2008). Such exhumation is also suggested by one AFT grain-age population of 27 Ma reported from a gneiss sample of the Dorud metamorphic complex of the Sanandaj–Sirjan Zone (Homke *et al.* 2010). Propagation of shortening in the Zagros Folded Belt and uplift associated with basement-involved thrusting did not occur before 12.4 Ma in the Fars region, thus placing constraints on the timing of plateau uplift.

4. Distribution of shortening and uplift in the Zagros, Iranian plateau and the Alborz

4.a. Distribution of shortening, underthrusting and underplating in the Zagros

The shortening within the Zagros belt appears highly inhomogeneously distributed between the Zagros Folded Belt to the south and the north where it is accommodated below the Sanandaj–Sirjan Zone (Figs 1a, 2). Among the total shortening accommodated in the Zagros belt, only 5 % (15 km) is taken up in the Zagros Folded Belt (Mouthereau *et al.* 2007). Next, I verify whether this value, obtained in the Fars province, is acceptable in the light of geophysical data and observed topography. Provided that the initial crustal thickness H_c is known and the amount of shortening (a − b), where a and b are the initial and the final lengths of the studied geological section, respectively, can be derived, the resulting Airy-compensated topography h is given by

$$h = \frac{(\mathrm{a}-\mathrm{b})H_c\Delta\rho}{\mathrm{b}\rho_m} \qquad (1)$$

where $\Delta\rho = \rho_m - \rho_c$ with $\rho_c = 2800$ kg/m^3 and $\rho_m = 3330$ kg/m^3.

In the first case, by assuming conservation of mass and in-plane deformation, and the fact that the related topographic load wavelength (i.e. 100 km) is too small with respect to the elastic thickness of the Arabian plate (T_e = 50 km; Snyder & Barazangi, 1986) to be compensated by a crustal root (Paul *et al.* 2006, 2010), the predicted topographic elevation of 2.25 km is simply obtained by equating initial and final crustal areas with H_c = 45 km. Even though a better result (i.e. elevation of 1.6 km) can be obtained for a lower shortening of 3 % (10 km), this calculation shows that only a small amount of shortening can account for the Zagros Folded Belt topography. In contrast, any greater shortening estimates would have resulted in unrealistic topographic elevations.

Northward, beneath the Sanandaj–Sirjan Zone, the shortening of the Arabian crust is seen to increase up to 37 % (50 km) and is thought to result from duplexing (Mouthereau *et al.* 2007). Prior to accretion of Arabian material below the Sanandaj–Sirjan Zone, during the early stages of the collision, the thinner and more distal portion of the Arabian margin was underthrusted. This is attested by receiver functions in the NW Zagros, revealing that the underthrusting of the Arabian crust below the obducted ophiolitic complex and Sanandaj–Sirjan Zone might have been as large as 250 km (Paul *et al.* 2010). However, only a part of it has been accommodated after Miocene time and hence can be considered in our calculation. Moreover, in the NW Zagros, Agard *et al.* (2005) showed that 50–70 km of Miocene shortening was taken up in the vicinity of, or at, the suture zone mainly within the ophiolitic sheets and thrust slices of the southern Sanandaj–Sirjan belt. These 50–70 km can represent 20–30 % of the total amount of shortening absorbed during the underthrusting of the Arabian margin as inferred from geophysics. As a result, they are not equivalent to the 37 % (50 km) of Mouthereau *et al.* (2007) accommodated by duplexing below the Main Zagros Thrust and instead must be added to them. One deduces that a total shortening of 135 km occurred near the suture zone and has likely been distributed as follows: 15 km in the Zagros Folded Belt (post-12.4 Ma), 50 km by duplexing (post-25 Ma) and up to 70 km by underthrusting (post-25 Ma) below the suture zone.

To explain this distribution I propose that the initial crustal configuration at 25 Ma, just before the initiation of thickening of the Arabian crust and its exhumation, resulted from the vertical stacking of three main units: (1) the thinned and flexed Arabian continental crust underthrusted below Central Iran by 50–70 km, (2) the overriding Neyriz ophiolitic complex made up of the oceanic lithospheric mantle emplaced in Late Cretaceous time and (3) the southern distal margin of the Eurasian continental crust corresponding to the Sanandaj–Sirjan Zone, which was essentially thickened during Jurassic and Early Cretaceous time.

To maintain a constant elevation of 2 km between the uncompensated Zagros Folded Belt and the adjacent domain of the suture zone exhibiting a crustal thickness of ∼ 70 km and shortening of 37 %, one should infer a denser crust (ρ_c = 3000 kg/m^3), likely related to the obducted mantle sheet. The predicted initial crustal thickness is of the order of 40–45 km, equivalent to the unthickened part of the Arabian margin (Gök *et al.* 2008). One can infer from these calculations that a simple assumption of inhomogeneously distributed in-plane shortening can explain the observed 25 km Moho deepening beneath the suture zone and the observed topography.

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4.b. Thickening of the Iranian plateau

To the north of the Sanandaj–Sirjan Zone, the mean Iranian plateau elevation is 1500 m according to Hatzfeld & Molnar (2010). Assuming that shortening occurred through Airy compensation, these authors estimated using the same equation in the previous Section that the crustal root would be 10–12 km to maintain the current topography. They derive an initial crustal thickness of 35–40 km. In an alternative view, they considered that the topography is not fully compensated by a buoyant crustal root but that at least 500 m could be accounted for by mantle delamination beneath the Iranian plateau.

One available estimate of shortening in Central Iran, north of the Urumieh–Dokhtar Magmatic Arc, is 38 km (29 %) and is thought to have occurred since 10 Ma (Morley *et al.* 2009). The current crustal thickness beneath Central Iran, also called Central Domain (CD) in Paul *et al.* (2010), is ~ 42 km or 48 km close to Alborz according to Radjaee *et al.* (2010). Assuming Airy compensation, the ~ 1 km elevation implies a crustal root of only 5 km, thus suggesting limited crustal shortening of only 14 %, which is significantly smaller than the value obtained from the balanced cross-section. Reconciling the observed shortening with the current crustal thickness and elevation requires increasing the average density of the Iranian crust to $\rho_c = 3000$ kg/m^3. This could be justified if the average composition of the Iranian crust has been substantially modified by magmatic underplating or by Eocene magmatic intrusions well described in the region (e.g. Allen & Armstrong, 2008). An average initial thickness of 32 ± 2 km is obtained. The geological meaning of the crustal thinning is probably two-fold. First, the development of Eocene deep-water basins to the north of the Urumieh–Dokhtar volcanic arc has been already noticed (Vincent *et al.* 2005 and references therein) and might be related to the regional back-arc extension episode (Vincent *et al.* 2005; Verdel *et al.* 2007; Morley *et al.* 2009). Second, a renewed episode of extension during Late Miocene time of unclear geodynamic origin (Morley *et al.* 2009) surely contributed to the crustal thinning. Finally, given the proposed 29 % of shortening over the entire length of the Iranian plateau (300–450 km), a shortening of ~ 120–180 km is obtained to build the current crustal thickness.

4.c. Timing and amount of shortening in the Alborz and the Caspian Sea

Shortening across the Alborz is estimated to range between 30 and 56 km (Allen *et al.* 2003; Guest *et al.* 2006*a*) and probably began between ~ 17 Ma, if one considers the increase in accumulation rates (Ballato *et al.* 2008, 2011), and 12 Ma ago (Guest *et al.* 2006*b*) in the Western Alborz or 6–4 Ma in the Central Alborz (Axen *et al.* 2001) if rapid exhumation is taken into account (Fig. 1b). Shortening associated with the subduction of the Caspian Sea to the north beneath the Apsheron Sill is constrained by the depths of earthquakes of at least 80 km (Jackson *et al.* 2002). Considering uncertainties in the timing of subduction initiation, I consider a value of ~ 75 km to be accommodated within this region, thus satisfying both the data and the total convergence of 440 km (Fig. 1a, b).

5. Discussion and conclusions

The absence of change in Arabian plate motion since 22 Ma (ArRajehi *et al.* 2010) just after the decrease from 3 to 2 cm yr^{-1} caused by the initiation of crustal thickening in the Zagros implies 440 km of Arabia–Eurasia convergence. This was accommodated since Miocene time across the Zagros belt, Central Iran, the Alborz and the Caspian Sea but not necessarily at the same rate. By taking into consideration the published amounts of long-term shortening and their timing, I suggest that it is possible to reproduce the total convergence predicted by geodetic and plate reconstruction (Fig. 7). If one refers to Figure 2, which is based on the balanced cross-section by Mouthereau *et al.* (2007) of the Fars arc region and on the study by Agard *et al.* (2005) to the north of the Lorestan arc region, about 135 km of convergence has been accommodated by frontal accretion in the Zagros Folded Belt (15 km), by duplexing (underplating) of Arabian crust below the Sanandaj–Sirjan Zone (~ 50 km) and by underthrusting (~ 70 km) localized across the Main Zagros Thrust. A maximum shortening of 180 km is obtained if in-plane shortening of 29 % is assumed to have occurred throughout Central Iran; 50 km were accommodated across the Alborz and 75 km were taken up by subduction of the Caspian Sea.

Thermochronologic data and age constraints on the initiation of the siliciclastic sedimentation in the foreland basins reveal that deformation initially concentrated in the Zagros *c.* 20 Ma (Homke *et al.* 2009; Gavillot *et al.* 2010; Khadivi *et al.* 2010; S. Khadivi, unpub. Ph.D. thesis, Univ. Pierre et Marie Curie, 2010) and in the Alborz approximately at the same time 20–17.5 Ma ago (Ballato *et al.* 2008, 2011) (Figs 1b, 7).

This stage was followed by propagation of shortening in the Zagros Folded Belt (Khadivi *et al.* 2010) and uplift in the Zagros after ~ 12.4 Ma (Figs 1, 7). This timing is concordant with the acceleration of deformation in the Alborz (Guest *et al.* 2006*b*), in the Kopet-Dagh and is coeval with the initiation of subduction of the south Caspian Sea (Hollingsworth *et al.* 2010) and deformation in Central Iran (Morley *et al.* 2009). Rapid exhumation in the Central Alborz at ~ 5 Ma (Axen *et al.* 2001) and coeval onset of increasing accumulation rates in the south Caspian Sea at 5.5 Ma (Allen *et al.* 2002), though possibly suggesting a younger subduction, also support the regional changes at 15–5 Ma (Figs 1b, 7).

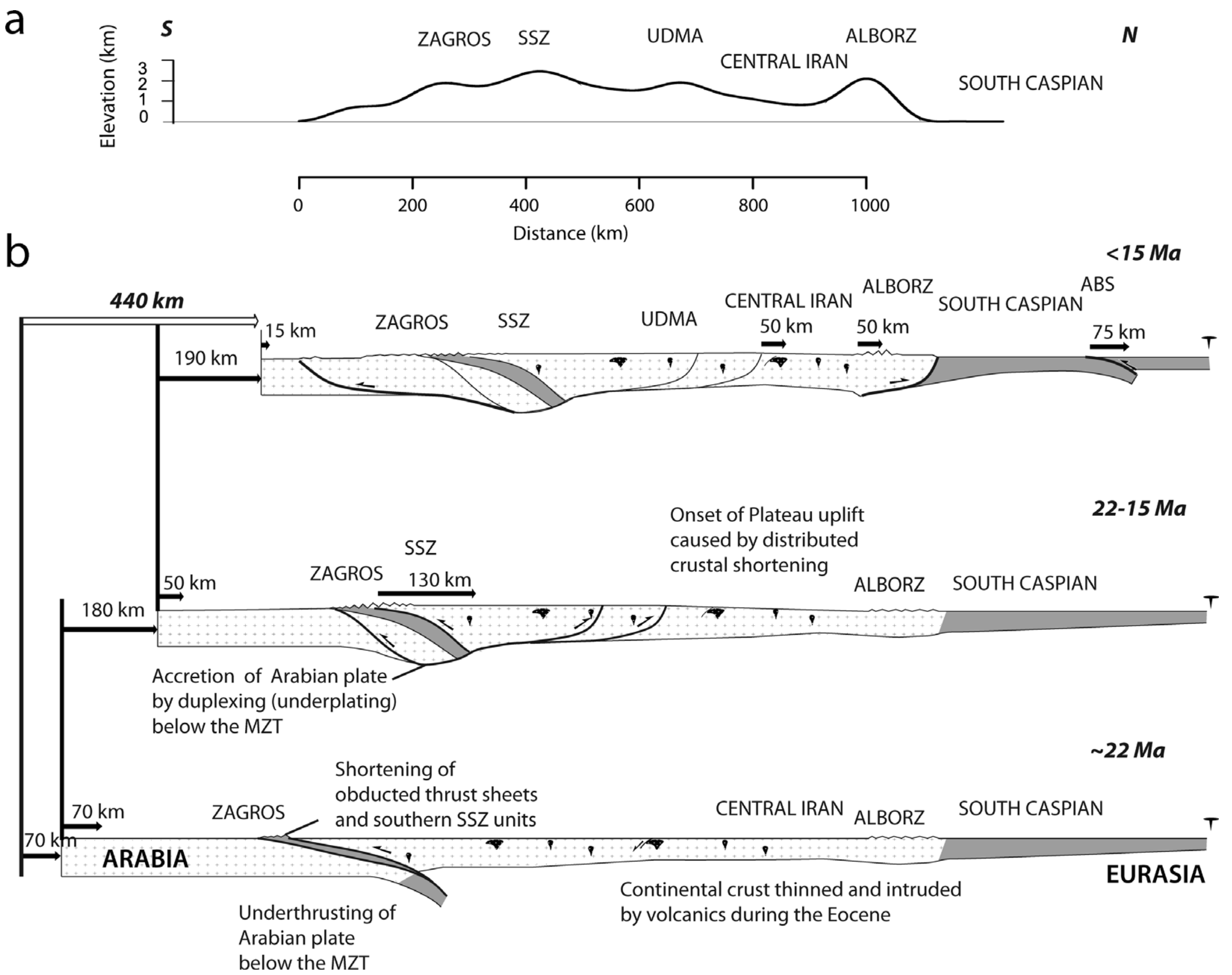

Figure 7. (a) Present-day topography and location of main tectonic belts in the Arabia–Eurasia collision for reference. (b) Distribution of shortening within orogenic belts and the Iranian plateau illustrating how the Arabian–Eurasian plate convergence was accommodated during the last 22 Ma. Note the progressive migration of shortening towards the north and in areas originally at low elevation. Abbreviations: SSZ – Sanandaj–Sirjan Zone; UDMA – Urumieh–Dokhtar Magmatic Arc; ABS – Apsheron–Balkan Sill; MZT – Main Zagros Thrust.

I propose that during the past 22 Ma stable motion of Arabia, a shift of localized deformation occurred in Late Miocene–Pliocene times toward the Zagros or the Alborz that were uplifting (Fig. 7). A concomitant decrease of shortening rates in the Iranian plateau occurred to compensate for constant boundary velocity. The insignificant change in Arabian plate motion makes the distribution of crustal shortening and underthrusting during the Arabia/Eurasia convergence the main driver of Zagros mountain and Iranian plateau uplift over the past 20 Ma. Slab detachment, which is suspected to be responsible for Miocene–Pliocene magmatic pulses, should therefore be considered with caution if we are to evaluate its contribution to the uplift of the whole Zagros region. I have herein proposed that the current topography of Central Iran can be explained by differences in the initial (i.e. before 20 Ma) thickness of the continental crust. This thinning of Central Iran is thought to be at least partly caused by a back-arc extensional regime related to the Neo-Tethyan slab rollback during Eocene time (Vincent *et al*. 2005; Moritz, Ghazban & Singer, 2006; Verdel *et al*. 2007; Morley *et al*. 2009; Ballato *et al*. 2011). The Iranian lithosphere was consequently relatively weak and hence shortened at low deviatoric stresses causing the inversion of extensional basins during Early Miocene time until its crust attained its present-day thickness. Because the crust of Central Iran became progressively thicker, the forces necessary to balance the increase of potential energy associated with plateau growth led to the reactivation of surrounding orogenic domains i.e. the Alborz and the Zagros after 12 Ma.

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