

## Review Article

# Building the Zagros collisional orogen: Timing, strain distribution and the dynamics of Arabia/Eurasia plate convergence

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## ABSTRACT

The Zagros Mountains are the result of the Arabia/Eurasia collision initiated at ~35 Ma as the rifted Arabian lithosphere was underthrust beneath the Iranian plate due to its negative buoyancy. The onset of crustal thickening started at ~25 Ma, as recorded by the hinterland exhumation and foreland clastic deposition. Deformation throughout the Arabia/Eurasia collision zone and the uplift of the Iranian plateau occurred after 15–12 Ma, as a result of shortening/thickening of the thin Iranian crust. We emphasize that only 42% of the post-35 Ma convergence is partitioned by shortening within central Iran. Tomographic constraints show ongoing slab steepening or breakoff in the NW Zagros, whereas underthrusting of the Arabian plate is observed beneath central Zagros. The current subduction dynamics can be explained by the original lateral difference in the buoyancy of the distal margin that promoted slab sinking in NW Zagros and underthrusting in central Zagros. Critical wedge approach applied to the Zagros favors the hypothesis of strong brittle crust detached above a viscous lower crust. In contrast, the weak sedimentary cover deforms by buckling of a thick multilayered cover. Thrust faulting associated with folding occurs in the competent layers and is responsible for most of the earthquakes. There is evidence that the role of the slab pull force in driving the Arabian plate motion was reduced after ~12 Ma. Large-scale mantle flow induced by mantle upwelling at the Afar plume appears to be the main driver of the Arabia plate motion. We stress that the main kinematic change in the Zagros region occurred at 15–12 Ma as the Zagros uplifted, before the Arabian slab detached. The Zagros appears key to investigate coupling between continental rheology, plate driving forces and mountain building, in which the role of rift inheritance appears to be central.

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## 1. Introduction

The Zagros is the largest mountain belt and the most active collisional orogen associated with Arabia/Eurasia convergence (Fig. 1). It belongs to the Alpine–Himalayan orogenic system that resulted from the closure of the Neotethys Ocean during the Cenozoic (Dercourt et al., 1986; Dewey et al., 1973; Stampfli and Borel, 2002). In this aspect, our current understanding of the Mesozoic geodynamics of the Zagros region, associated with the subduction of the Neotethys ocean, has been largely improved in the past few years thanks to the synthesis of an increasing number of geochronological constraints on tectono-magmatic events combined with tomography data (e.g., Agard et al., 2011; Verdel et al., 2011 for a review).

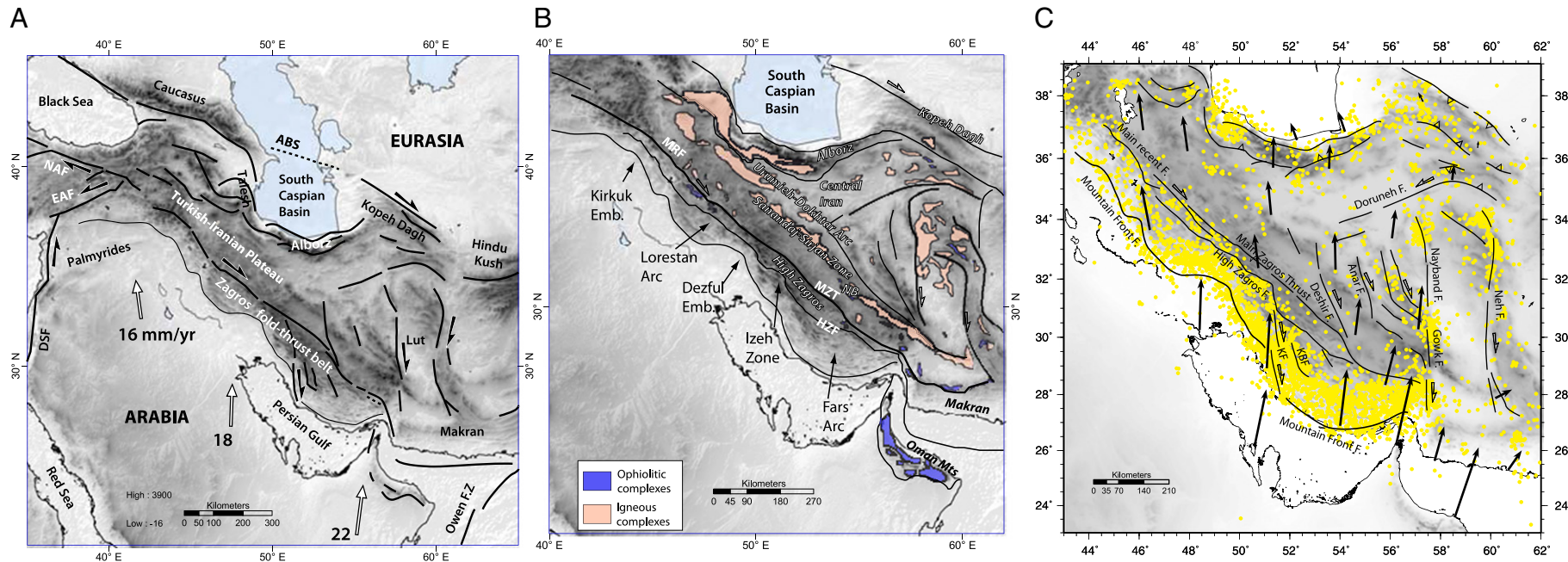
In contrast, the temporal evolution of the Zagros topography and adjacent Iranian plateau uplift is much less well understood. It is not clear whether the regional uplift was achieved rapidly in the past 10–5 Ma (e.g., Agard et al., 2005, 2011; Allen et al., 2004; Allen et al., 2011; Molinaro et al., 2005a,b) or was instead progressive in the whole collision region over the last 35 Myr (e.g., Ballato et al., 2011; Mouthereau, 2011). Also our understanding of the dynamics of mountain building mostly relies on a number of emergent tectonic models, not mutually exclusive, highlighting the role of distributed shortening in the Arabia/Eurasia collision, Arabian slab breakoff and mantle delamination (Agard et al., 2005; Ballato et al., 2011; Hatzfeld and Molnar, 2010; Molinaro et al., 2005b; Mouthereau, 2011; van Hunen and Allen, 2011).

Answering these questions is of main importance since the uplift of the Zagros region is thought to have influenced the connectivity between the Indo-Pacific Ocean, the Mediterranean Sea and the Tethys Ocean (Harzhauser et al., 2007; Kocsis et al., 2009; Reuter et al., 2009) and to have potentially impacted the Cenozoic global climate (Allen and Armstrong, 2008). Yet, our knowledge of temporal evolution of land/sea distribution in the Zagros region, including the

Iranian plateau, is still imprecise regardless of the recent efforts to build paleotectonic maps (Barrier and Vrielynck, 2008).

In the framework of the Africa/Eurasia convergence, different causes of the Arabian plate motion has been proposed, including slab pull effects or/and mantle drag (Alvarez, 2010; ArRajehi et al., 2010; Becker and Faccenna, 2011; Bellahsen et al., 2003; Faccenna et al., 2006; Hafkenscheid et al., 2006; Jolivet and Faccenna, 2000; McQuarrie et al., 2003; Reilinger and McClusky, 2011). A more accurate knowledge of the timing of uplift and distribution of crustal thickening at the Zagros collisional plate boundary is therefore needed to advance on this question.

A significant progress has been made in establishing a reliable stratigraphic framework, which allowed the timing of collision onset (Agard et al., 2005; Allen et al., 2004; Fakhari et al., 2008; Gavillot et al., 2010; Horton et al., 2008; Khadivi et al., 2012; Vergès et al., 2011b), the sequence of deformation (Ahmadhadi et al., 2007; Authemayou et al., 2005; Emami, 2008; Gavillot et al., 2010; Hessami et al., 2001; Homke et al., 2004; Khadivi et al., 2010; Molinaro et al., 2005a; Mouthereau et al., 2007b) and the chronology of cooling/denudational events (Homke et al., 2010; Khadivi et al., 2012) to be more precisely known. At shorter time scale, geodetic measurements of the Arabia plate motion combined with longer term fault slip rates deduced from seismotectonic studies of Quaternary and active faults, gave a more accurate picture of the way arc-normal and strike-slip components of the N-directed Arabia/Eurasia convergence is accommodated in the Zagros region (Allen et al., 2004; ArRajehi et al., 2010; Authemayou et al., 2006, 2009; Lacombe et al., 2007; McClusky et al., 2003; McQuarrie et al., 2003; Nilforoushan et al., 2003; Talebian and Jackson, 2004; Vernant et al., 2004; Walpersdorf et al., 2006). This led to propose a profound kinematic reorganization at 5–7 Ma (Allen et al., 2004), which is still challenged by plate kinematic data arguing, given uncertainties, a stable Arabia plate convergence since ~22 Ma (ArRajehi et al.,



**Fig. 1.** Main tectonic features and kinematics of the Arabia/Eurasia collision. A) Main morphotectonic units and active faults of the Arabia/Eurasia convergence draped onto shaded relief from ETOPO1 (1' x 1' resolution) Global Relief data (<http://www.ngdc.noaa.gov>). White arrows refer to the relative Arabian plate motion with respect to fixed Eurasian plate (Sella et al., 2002; Vernant et al., 2004). B) Main tectono-magmatic belts of the Zagros, arcs and recesses at the Zagros front and location of obducted ophiolites. C) Active faults, GPS velocities shown as black arrows are from Masson et al. (2007) and the 1964–2002 seismicity from International Seismological Centre (2001). Abbreviations are Main Zagros Thrust (MZT), High Zagros (HZ), Apsheron-Balkan Sill (ABS), East Anatolian Fault (EAF), North Anatolian Fault (NAF), Dead Sea Fault (DSF), Main Recent Fault (MRF), High Zagros Fault (HZF), Mountain Front Fault (MFF), Kermanshah ophiolite Complex (K), and Neyriz ophiolite Complex (N).

2010). However, these data have never been integrated to propose a comprehensive scheme of the strain distribution over the whole collisional plate boundary.

The Zagros foreland fold-thrust belt is particularly famous as the most prolific fold-thrust belt where the world's largest hydrocarbon reserves are trapped in giant anticlines (e.g., Bordenave and Hegre, 2010). In addition to exceptional outcrop conditions, this has made the Zagros fold-thrust belt one fascinating area for studying the mechanics of fold development (Blanc et al., 2003; Casciello et al., 2009; Lacombe et al., 2007; McQuarrie, 2004; Mitra, 2003; Molinaro et al., 2005a; Mouthereau et al., 2006, 2007a, 2007b; Oveisi et al., 2009; Sherhati et al., 2005; Vergés et al., 2011a; Yamato et al., 2011) and the interplays between folding, sediment routing and landscape evolution (Khadivi et al., 2012; Oberlander, 1985; Ramsey et al., 2008; Tucker and Slingerland, 1996).

Past structural interpretations of the Zagros structure emphasized the along-strike variability of the structural styles, including thin-skinned deformation and basement-involved thrusting. In this framework, the specific role of weak detachment levels in the cover and the rheology of the basement in building the Zagros topography has been questioned (Bahroudi and Koyi, 2003; Dahlen, 1990; Davis and Engelder, 1985; Ford, 2004; Mouthereau et al., 2006). To advance on these questions, a better understanding of the mechanical coupling between the rheology and tectonic inheritance on the Arabian margin, plate kinematics, and change in boundary forces is needed here.

The goal of this review paper is to investigate the dynamic evolution of the Zagros orogenic system during the Cenozoic, following a very active decade of acquisition of geologic, geochronological and geophysical datasets. These local to regional scale data are integrated together with other constraints from tectonic belts of central, western and eastern Iran in order to examine the dynamic coupling between plate convergence and mountain building in the Zagros/Iranian Plateau. This review study provides new keys for a better understanding of the driving forces at the origin of the Zagros orogeny and constraints on the paleogeographic evolution of this key region between the Mediterranean Sea and Indian–Pacific ocean world. We finally propose a new geological and geodynamic evolutionary model of the Zagros collision since 55 Ma that accounts for the progressive strain distribution and exhumation/uplift of the Zagros collision. This model provides new view of the role of the initial margin geometry in governing the distribution of underthrusting and accretion in the collision domain.

## 2. Geological setting

The Zagros mountain belt is defined as a NW-trending orogen stretching 2000 km from the East Anatolian fault in eastern Turkey (45°E, 36°E) to the Makran subduction in southern Iran (26°N, 58°E). The Zagros orogen is flanking the Turkish–Iranian plateau to the south and its elevation reaches a maximum of 4548 m in the Khuzestan province of the NW Zagros (Fig. 1). In Iran, the Mesozoic and Cenozoic convergence between Africa/Arabia–Eurasia resulted in NW-trending parallel tectono-metamorphic and magmatic belts. These are the Zagros (or Simply) Folded Belt, the Imbricate Zone including the Kermanshah and Neyriz ophiolitic complexes, the Sanandaj–Sirjan Zone and the Urumieh–Dokhtar volcanic arc (Fig. 1)

Berberian and Berberian, 1981; Berberian and King, 1981; Berberian et al., 1982).

### 2.1. Zagros Folded Belt (ZFB): structural segmentation, folding and faulting

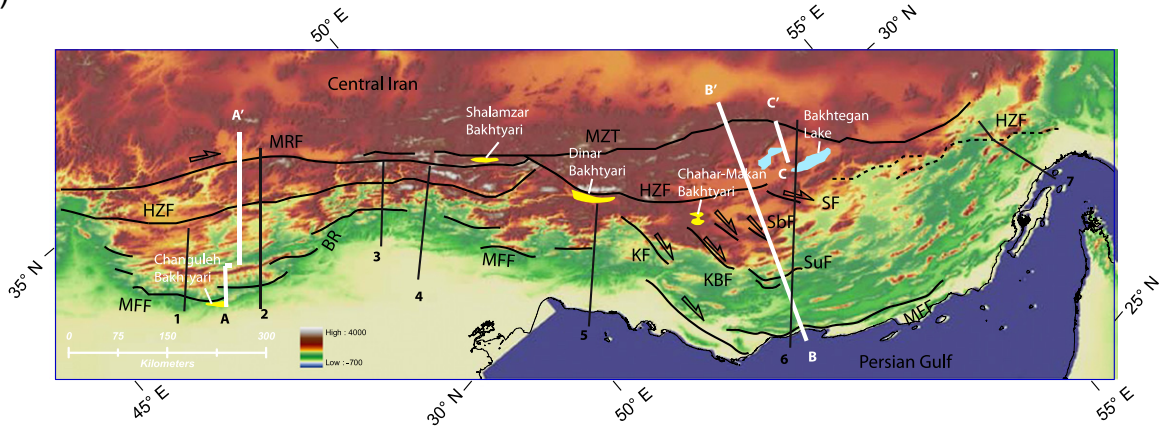
The ZFB is separated into several domains according to the along-strike changes in structural styles, position of the deformation front and stratigraphy (Fig. 1b). These changes reflect a kinematic segmentation along N-trending faults inherited from the Late Proterozoic fault system of the Pan-African basement (Talbot and Alavi, 1996). Several of these inherited features are recognized in the Arabian Shield and can be followed northwards, into the Zagros Basin before they were reactivated in the Zagros collision belt (Bahroudi and Talbot, 2003; Berberian, 1995; Hessami et al., 2001; Talbot and Alavi, 1996). These domains are from North to South the Kirkuk embayment, the Lorestan (Pusht-e Kuh arc), the Dezful embayment, the Khuzestan recess (Izeh Zone) and the Fars arc (Casciello et al., 2009; Falcon, 1974b; Lacombe et al., 2006; Sherhati and Letouzey, 2004; Stocklin, 1968). The Dezful embayment is bounded by the Dezful Embayment Fault (DEF) to the north, the Balarud Line to the west, the Kazerun Fault to the east (Sepehr and Cosgrove, 2004), the Mountain Front Fault (MFF) to the northwest and the Zagros Foredeep Fault (ZFF) to the southwest. The Izeh domain is delimited to the north-northeast by the High Zagros Fault (HZF) and to the south-southwest by the Mountain Front Fault (MFF). Its northwestern limit lies along Balarud Line. The Fars arc stretches 300 km and is delineated by the MFF to the south, the Kazerun Fault to the west and by the Minab–Zendan fault system to the southeast, which outlines the active fault boundary between the Zagros collision and the Makran subduction (e.g., Regard et al., 2004, 2005).

The Zagros folds formed in a thick pile of sedimentary rocks up to 12 km (Colman-Sadd, 1978; Falcon, 1974a; James and Wynd, 1965; Stocklin, 1968) including Paleozoic, Mesozoic and Cenozoic strata (Figs. 2 and 3). These were deposited in an extensional and passive margin setting during Paleozoic and most of the Mesozoic periods followed by compression and flexural basin development starting in the Late Cretaceous times (e.g., Beydoun et al., 1992; Homke et al., 2009 and Koop and Stoneley, 1982 among others). Tertiary foreland sequences are represented by the Fars Group comprising the Gasharan/Razak, Mishan and Agha Jari formations, which altogether forms a regressive siliciclastic sequence of ~3 km, overlying the carbonate platform of the Oligo-Miocene Shahbazan and Asmari formations, the latter being the main Tertiary oil reservoir.

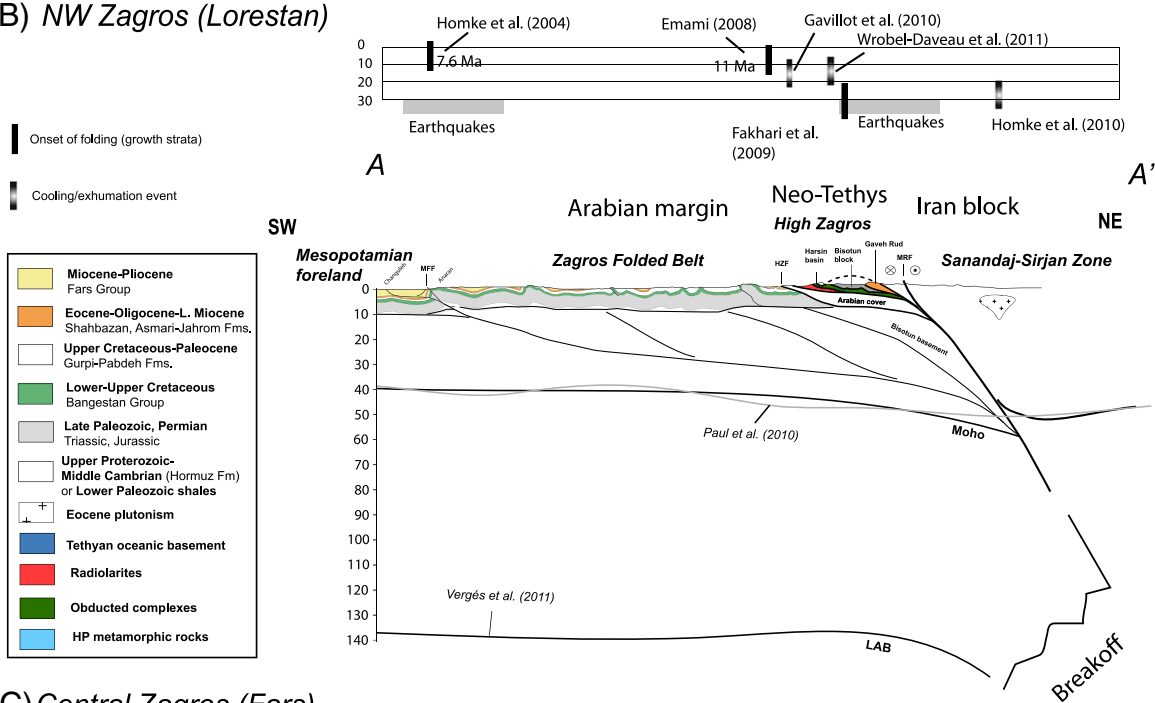
In the Fars, the base of the sedimentary cover rocks overlies the infracambrian salt (Hormuz Formation), which maximum thickness is 1–2 km and which acts as an extremely efficient décollement level (Colman-Sadd, 1978; Edgell, 1996). Together with the Miocene evaporitic formations, these units are known to be particularly mobile and forms one of the largest fields of salt diapirs worldwide, most of them being concentrated in the eastern Fars arc (e.g., Jahani et al., 2009; Talbot and Alavi, 1996). Several Hormuz salt plugs are cropping out along the southern faulted boundary of the High Zagros (High Zagros Fault, Figs. 1 and 2) and N-trending strike-slip segments of the ZFB, constraining the minimum extent of this unit at depth (Fig. 2a). The lack of diapirs and plugs in the Dezful embayment and Lorestan arc has been attributed to the absence of Hormuz salt in

**Fig. 2.** Crustal-scale structure and temporal constraints on fold/thrust emplacement in NW Zagros and central Zagros. A) Location of balanced cross-sections in the Zagros: 1) Alavi (2007); 2) McQuarrie (2004); 3) Blanc et al. (2003); 4) McQuarrie (2004); 5) Sherhati and Letouzey (2004); 6) McQuarrie (2004); and 7) Molinaro et al. (2005a). Sections AA', BB' and CC' correspond to crustal balanced cross-sections of the Lorestan and the Fars regions after Vergés et al. (2011b) and Mouthereau (2011), respectively. Position of the Bakhtyari conglomerates dated by magnetostratigraphy for the Changuleh Bakhtyari (Homke et al., 2004), by biostratigraphy for Shalamzar Bakhtyari (Fakhari et al., 2008), indirectly by low-temperature thermochronometry in the Dinar Bakhtyari dated (Gavillot et al., 2010) and by magnetostratigraphy and biostratigraphy for the Chahar-Makan Bakhtyari (Khadivi et al., 2010). Abbreviations: Mountain Front Fault (MFF), High Zagros Fault (HZF), Main Recent Fault (MRF), Main Zagros Thrust (MZT), Kazerun Fault (KF), Karebass Fault (KBF), Sabz-Pushan Fault (SBF), Surmeh Fault (SuF), Sarvestan Fault (SF), and Balarud Fault (BR). B) Lithospheric scale cross-sections of the NW Zagros (Lorestan) after area balanced cross-section proposed by Vergés et al. (2011b) and temporal constraints on fold/thrust emplacement. C) Lithospheric scale cross-section modified after balanced cross-section after Mouthereau et al. (2007b) and Mouthereau (2011) for the Fars region. Section CC' is projected.

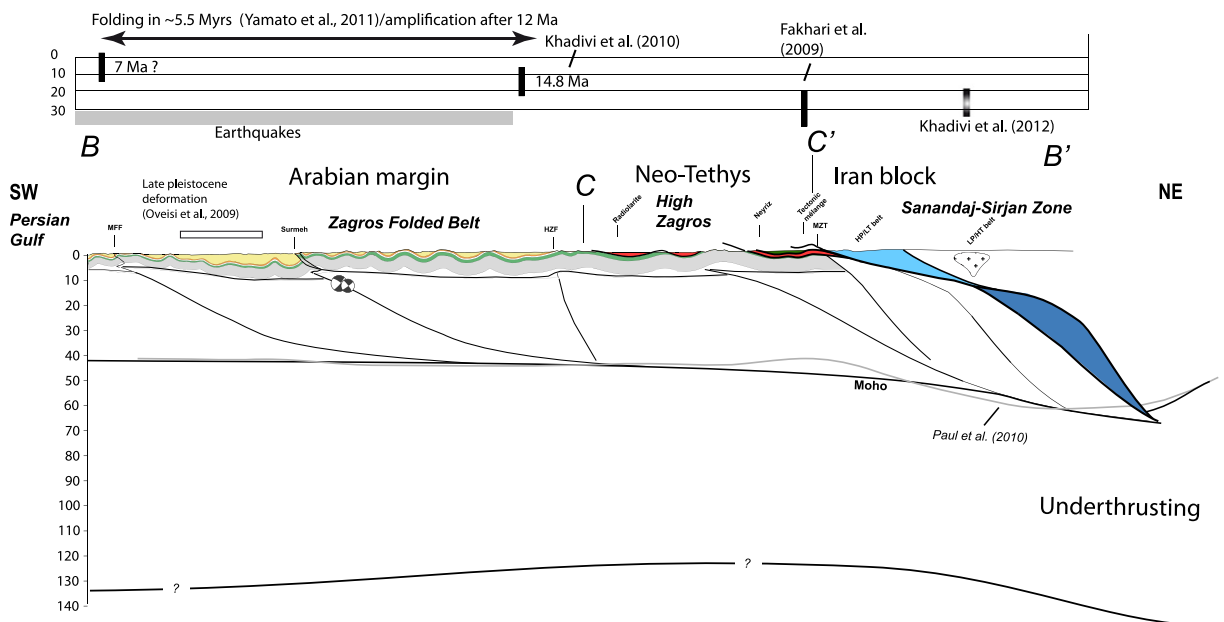
A)



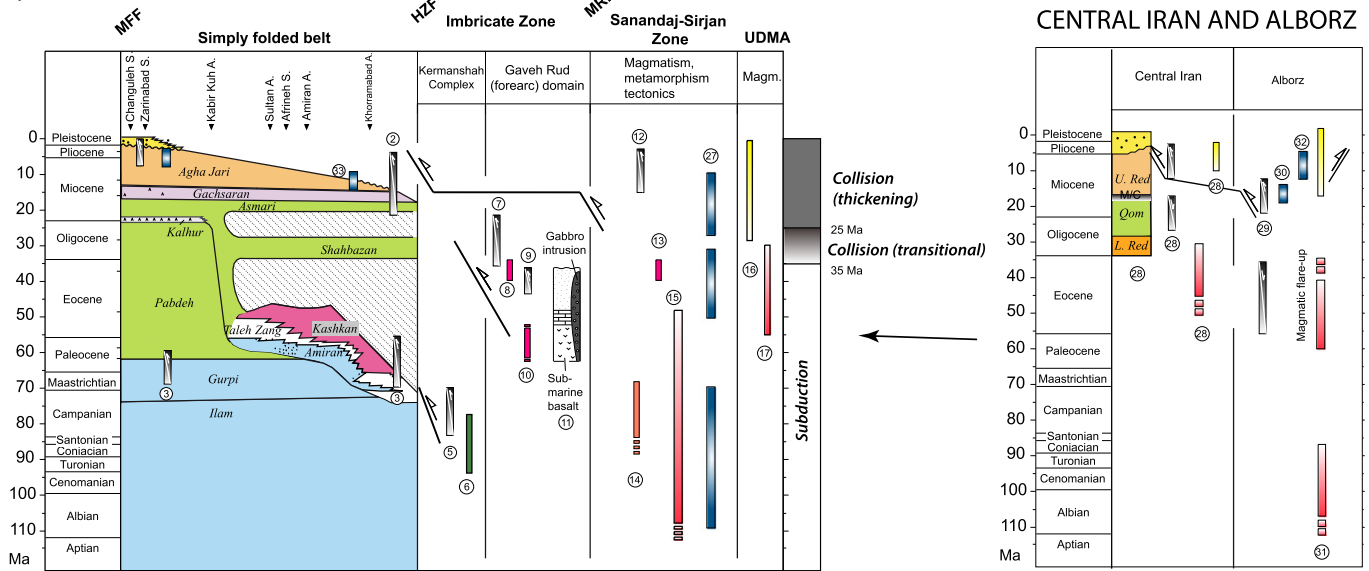
B) NW Zagros (Lorestan)



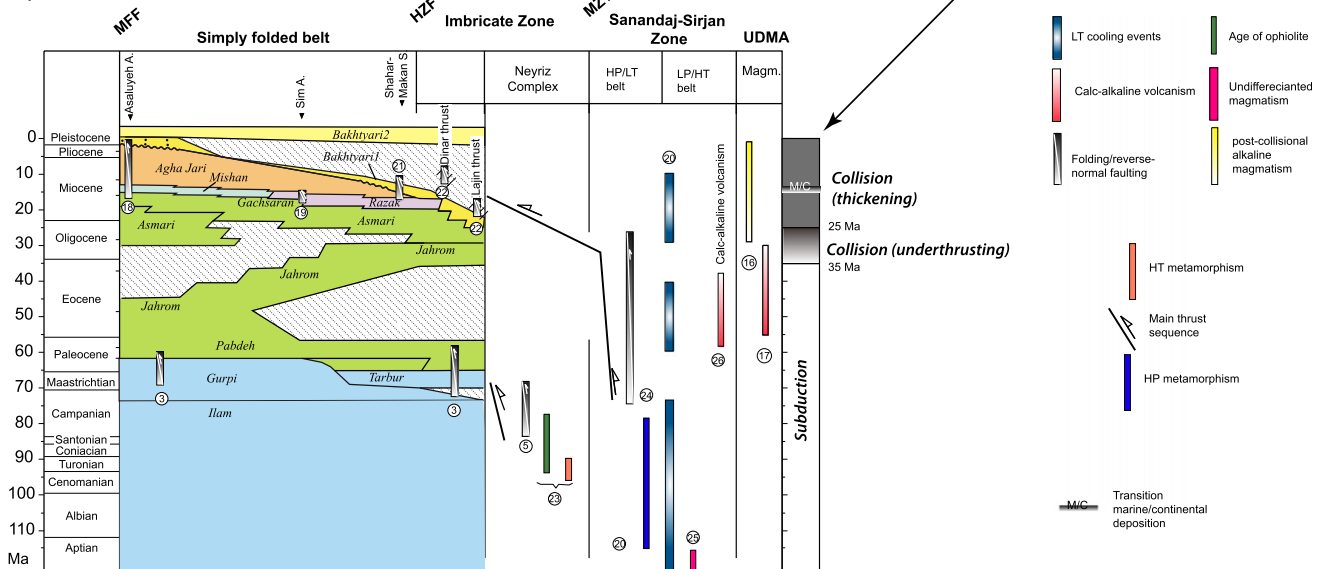
C) Central Zagros (Fars)



## A) LORESTAN (NW Zagros)



## B) FARS (Central Zagros)



**Fig. 3.** Synthesis of basin stratigraphy and compilation of thermochronological/geochronological constraints on tectonic, metamorphic, magmatic and exhumational events in the NW Zagros (Lorestan) and Central Zagros (Fars) organized by structural zones and in central Iran and the Alborz. Kermanshah Complex, Kermanshah radiolarite–ophiolite Complex; Neyriz Complex, Neyriz radiolarite–ophiolite Complex; MFF, Mountain Front Fault; HZF, High Zagros Fault; MZT, Main Zagros Thrust; MRF, Main Recent Fault; and UDMA, Urumieh–Dokhtar Magmatic Arc. Compiled data are from (1) Homke et al. (2004); (2) Emami (2008), Fakhari et al. (2008); (3) Hessami et al. (2001), Sherkati and Letouzey (2004), Fakhari et al. (2008), Homke et al. (2009), (4) James and Wynd (1965), Gidon et al. (1974), Braud (1987), Homke et al. (2004), Agard et al. (2005), Homke et al. (2009), Khadivi et al. (2010); (5) Berthier (1974), Gidon et al. (1974), Braud (1987); (6) Delaloye and Desmons (1980); (7) Braud (1987), Agard et al. (2005), (8) Leterrier (1985) and Braud (1987); (9–11) Braud (1987), (12) Braud (1970), Gidon et al. (1974), Braud (1987), Agard et al. (2005); Fakhari et al. (2008); Khadivi et al. (2012), (13) Leterrier (1985), Braud (1987); (14) Mohajjel and Fergusson (2000), (15) Valizadeh and Cantagrel (1975), Berberian and Berberian (1981), Berberian et al. (1982), Masoudi (1997), Masoudi et al. (2002), Horton et al. (2008); (16–17) Bernard et al. (1979), Martel–Jentin et al. (1979), Berberian and Berberian (1981), Berberian et al. (1982), Bina et al. (1986), Aghazadeh et al. (2011); (18–19) Mouthereau et al. (2006), Mouthereau et al. (2007b); (20–21) Khadivi et al. (2010), Khadivi et al. (2012); (22) Gavillot et al. (2010); (23) James and Wynd (1965) Hallam (1976), Ricou (1976), Lanphere and Pamic (1983), Babaie et al. (2006), (24) Haynes and Reynolds (1980), Sarkarinejad and Alizadeh (2009), Sarkarinejad et al. (2009), (25) Berberian and Berberian (1981), Sheikholeslami et al. (2008), (26) Rachidnejad–Omran et al. (2002); Vincent et al. (2005), Moritz et al. (2006), Allen and Armstrong (2008); Morley et al. (2009), Ballato et al. (2011); (27) Homke et al. (2010); (28) Morley et al. (2009); (29) Ballato et al. (2008, 2011); (30) Ballato (pers. com.); (31) Guest et al. (2006a,b), (32) Axen et al. (2001) and Guest et al. (2006b); and (33) Wrobel–Daveau et al. (2011).

these two areas (Bahroudi and Koyi, 2003; Talbot and Alavi, 1996). The regular distribution and trend of the Zagros folds, however, strongly suggest they nevertheless formed as detachment folds above a weak basal layer equivalent to Cambrian shales, as proposed by Sherkati and Letouzey (2004) and discussed in Vergés et al. (2011a).

Fig. 2 shows two balanced crustal-scale cross-sections of the Zagros along which the crustal thickness is constrained by

geophysical data. The first one is located in the NW Zagros (Lorestan region from Vergés et al., 2011b) and the second in the central Zagros (Fars region modified from Mouthereau et al., 2007b). In both sections, the folded sediment cover is detached in the Hormuz salt or Lower Paleozoic shales over the Precambrian basement. The basement is also involved in shortening but its importance is debated. Yet controversial, the idea of involving the Precambrian basement in

deformation is supported by past and recent seismotectonic studies (Berberian, 1995; Jackson, 1980; Ni and Barazangi, 1986; Roustaei et al., 2010; Talebian and Jackson, 2004; Tatar et al., 2004). Among the active reverse faults reported from modern and historical earthquakes (e.g., Berberian, 1995) are the Mountain Front Fault (MFF; Figs. 1c and 2), the Dezful Embayment fault (DEF), and the Zagros Foredeep fault (ZFF). Re-appraisal of fault plane solutions for earthquakes along these reverse faults indicate that they dip 30–60° NE (Talebian and Jackson, 2004), suggesting they are inverted normal faults that originally developed during rifting of the Arabian continent (e.g., Berberian, 1995; Jackson, 1980). In a more recent study, Nissen et al. (2011) re-appraised the inferred basement earthquakes and relocated most of them in the cover, suggesting that the basement deforms by aseismic creep beneath the ZFB. They also noticed that some few large earthquakes (e.g., Mw~6.7) can rupture the basement.

Basement-involved shortening is also indicated by morphostructural observations, including abrupt changes in stratigraphic relief of the sedimentary cover across inferred deep active reverse faults (Berberian, 1995; Blanc et al., 2003; Emami et al., 2010; Lacombe et al., 2006; Molinaro et al., 2004; Mouthereau et al., 2007b; Sherkati and Letouzey, 2004). In addition, the sedimentary cover is cut by major NW-trending seismogenic strike-slip fault zones reported as active basement faults like the Surmeh–Karebass and Sabz–Pushan transverse fault zones (Figs. 1 and 2a). The Balarud Fault (BR, Fig. 2a) is one unique example of E–W active fault zone, bounding the Dezful Embayment to the north. Although, the Balarud seismic line has been interpreted as a major left-lateral strike-slip fault (Berberian, 1995; Hessami et al., 2001), focal mechanisms of earthquakes indicate thrust rupture within the basement and the cover (Allen and Talebian, 2011; Talebian and Jackson, 2004).

## 2.2. High Zagros and Zagros suture

### 2.2.1. Stratigraphic and tectonic features

The High Zagros (HZ) or Imbricate Zone (IZ) is characterized by the emplacement of imbricate thrust sheets. In the NW Zagros, to the north of the Lorestan, Dezful and Izeh Zone, several thrust sheets expose Palaeozoic strata, indicating the accommodation of large displacements at the origin of the highest topography in the region (Fig. 2). This contrasts with the central Zagros where HZ deformation is taken up by folding with large wavelengths.

The High Zagros is bounded to the north by the Main Zagros Thrust (MZT) also named the Main Zagros Reverse Fault (MZRF) or Main Zagros Fault (MZF), a north steeply-dipping thrust, which is connected southwards to a hangingwall flat carrying ultramafic suites of the Kermanshah and Neyriz obducted complexes. It is thought to represent the plate boundary between the folded cover of the Arabia margin to the SW and the upper magmatic–metamorphic belt of central Iran to the NE (Berberian and King, 1981) but this interpretation has been challenged (see discussion on the suture zone below).

In the NW Zagros (Fig. 2) the MZT is cut by the Main Recent Fault (MRF) a major active strike-slip fault partitioning the N–S convergence into a NW–SE right-lateral strike-slip and NE–SW shortening (Talebian and Jackson, 2002). The MZT appears to be mostly inactive as it is suggested by the absence of related seismicity. This is further indicated by its surface trace that rotates clockwise in the vicinity of the southern termination of the N-trending Deshir strike-slip fault, consistently with its right-lateral motion (Meyer et al., 2005).

To the south, the High Zagros is separated from the ZFB by the High Zagros Fault (HZF). As noticed by Talebian and Jackson (2004), most of the larger earthquakes occur in the ZFB (Fig. 1c) or along the High Zagros Fault (HZF) keeping the High Zagros devoid of seismicity, in agreement with the very few evidence from geodetic shortening (Tatar et al., 2002). Aseismic deformation indicates that shortening occurs either within ductile duplexes (e.g., beneath the

Sanandaj–Srijan zone) or has been entirely transferred to the south in the ZFB. While the HZF shows neither significant displacement nor remarkable geomorphic features in the Fars area, it accommodates large displacement in the NW Zagros as inferred from Cambrian rocks overthrusting the Cenozoic strata (Gavillot et al., 2010). There the HZF is displaying at least 6 km of vertical structural offset and is associated with historical and instrumental earthquakes in SE Zagros (Berberian, 1995).

The displacement accommodated along the MRF is currently being transferred in the central Zagros in N–S active right-lateral transpressional faults (Kazerun, Karebass, Sabz–Pushan, Sarvestan). These faults are partitioning longitudinally the oblique and frontal convergence between the NW and central Zagros, respectively (Authemayou et al., 2005, 2006, 2009; Lacombe et al., 2006; Talebian and Jackson, 2004). This kinematic change requires extension along the strike of the belt, which can be achieved progressively by the Kazerun/Karebass/Sabz–Pushan/Sarvestan faults if they rotate anticlockwise about vertical axes (Talebian and Jackson, 2004). Based on dating of geomorphic features, these faults have been alternatively interpreted as an orogen-scale strike-slip fault termination, which transfers and distributes orogen-parallel dextral slip, achieved along the MRF, into thrusts and folds of the Zagros belt (Authemayou et al., 2006, 2009).

### 2.2.2. The Zagros suture: Neyriz and Kermanshah obducted ophiolites

The Zagros suture zone (or Crush Zone) shows the imbrications of thrust units of different tectonic origins including rifted continental blocks, ophiolites and tectonic mélange (Fig. 2). The suture zone was emplaced as two obducted ophiolitic complexes next to the Fars (Neyriz) and the Lorestan region (Kermanshah) (Haynes and McQuillan, 1974; Stocklin, 1968) (Figs. 1b and 2b, c). Their distinctive distribution likely reflects their original position on an irregular continental margin (Allen and Talebian, 2011). The ophiolitic complex of Neyriz is considered to be an allochthonous fragment of the western Neo-Tethyan oceanic lithosphere (Golonka, 2004; Stocklin, 1968). It consists of several thrust imbricates. The uppermost tectonic unit includes gabbros, diorites, plagiogranites and variably serpentized peridotites (Babaie et al., 2006) and constitutes the crustal sequence of the Neyriz ophiolite well-exposed in Tang-e-Hana. East of lake Bakhtegan, this unit is overthrust by the Hajiabad mélange of upper Cretaceous age, which is composed of Permian–Triassic limestones, radiolarian cherts, tuffs, basalts (pillow lavas) and formed originally in a forearc setting (Babaei et al., 2005). Greenschist-to-amphibolite metamorphic rocks are found above the basal detachment shear zone of the allochthonous ophiolite complex (Babaie et al., 2006; Sarkarinejad et al., 2009). To the west of Lake Bakhtegan, both the Neyriz tectonic mélange and the ophiolite (Fig. 2) are thrustured over a sedimentary assemblage of radiolarian cherts, turbidites, middle Jurassic oolitic, micro-brecciated limestones, and the highly folded Pichakun formation dated from Late Triassic to Middle Cretaceous (Ricou, 1976; Robin et al., 2010).

The basal contact of the Neyriz obducted complex represents the hangingwall flat of the Main Zagros Thrust. The ophiolite complex was tectonically emplaced onto the Cenomanian–Turonian shallow-marine Sarvak Formation (Hallam, 1976) (Figs. 2 and 3).  $^{40}\text{Ar}/^{39}\text{Ar}$  dating on hornblende in diabase and plagiogranite yielded an age of 92–93 Ma (Babaie et al., 2006) consistent with ages of ~95 Ma obtained in amphibolites and slightly younger ages of ~86 Ma in tholeiitic sheeted dykes (Lanphere and Pamic, 1983). Together with the age of the unconformably overlying limestones of the Tarbur Formation, the ophiolites have therefore been emplaced between 86 Ma and 70 Ma (Fig. 3). Subsequent Zagros orogeny folded the originally flat thrust contact as illustrated by its contorted cartographic trace.

The Neo-Tethyan ocean origin for the Neyriz ophiolites has been widely proposed (Hallam, 1976; Haynes and Reynolds, 1980; Lanphere and Pamic, 1983; Stocklin, 1974). Geochemical studies

have pointed out that the obducted ophiolites may be originated from a mid-ocean ridge or Ca/K island arc (e.g., Hassanabad unit) (Babaie et al., 2001; Babaie et al., 2006), thus requiring an intraoceanic subduction (Agard et al., 2011; Arfania and Shahriari, 2009). It has been suggested that they were originated in a Red Sea-type rift in a shallow passive continental margin (Arvin, 1982; Stoneley, 1981). More recently, it has been proposed that they are suprasubduction ophiolites that resulted from the emplacement of an outer forearc oceanic basement, on the southern edge of the SSZ, formed when the subduction initiated in the mid-Cretaceous (Moghadam and Stern, 2011; Moghadam et al., 2010).

The Kermanshah ophiolitic complex includes 1) the folded and faulted Bisotun unit, composed of limestones of Triassic–Cenomanian age, 2) the Harsin colored mélange composed of upper mantle serpentinites, Eocene radiolarites, lavas and carbonates blocks, and 3) radiolaritic nappes which are dated from Lower Jurassic (Pliensbachian) to Upper Cretaceous (Turonian) (e.g., Wrobel-Daveau et al., 2010). Likewise Neyriz ophiolites, the Kermanshah ophiolitic complex was emplaced in the upper Cretaceous and subsequently thrust during the Zagros collisional orogeny in the Miocene. Wrobel-Daveau et al. (2010) re-appraised the ophiolitic suites of the Kermanshah Complex as upper mantle serpentinitized peridotites exhumed in the stretched Arabian margin. In this interpretation, the ophiolites are located between the radiolaritic basin and the rifted Bisotun continental block. This interpretation emphasizes the important differences in the pre-collisional margin configuration between NW and central Zagros. The implications for the geodynamics of the region and mountain building of the Zagros will be discussed later.

### 2.3. The Sanandaj–Sirjan metamorphic belt or Sanandaj–Sirjan Zone (SSZ)

The Sanandaj–Sirjan Zone (SSZ), located to the north of the MZT, represents the tectono-magmatic and metamorphic part of the Zagros belt (Figs. 1b, 2b, c and 3). It is made up of sedimentary and metamorphic Paleozoic to Cretaceous rocks formed in the former active margin of an Iranian microcontinent drifted during the Late Jurassic (Berberian and Berberian, 1981; Dercourt et al., 1986; Golonka, 2004), which collided with the Arabian margin during the Miocene. Alternatively, SSZ has been seen as the metamorphic core of the Zagros accretionary complex built by the thickening of distal crustal domains of the Arabian margin (Alavi, 2004; Moghadam et al., 2010). During the Middle Jurassic and Lower Cretaceous times, the SSZ was an active Andean-like margin characterized by calc-alkaline magmatic activity in which andesitic and gabbroic intrusions were emplaced (Agard et al., 2005; Berberian and Berberian, 1981). The subduction beneath SSZ have initiated in the Triassic (Berberian and Berberian, 1981; Dercourt et al., 1986; Kazmin et al., 1986) and possibly lasted until the Eocene (Mazhari et al., 2009). By mid-Cretaceous, magmatism shifted inland that is in the UDMA and the Alborz (Agard et al., 2011; Berberian et al., 1982; Kazmin et al., 1986; Verdel et al., 2011).

The metamorphic part of the Sanandaj–Sirjan Zone can be subdivided into HP/LT and HT/LP metamorphic belts that developed at a transpressional plate boundary associated with the subduction of Neo-Tethys (Sarkarinejad and Azizi, 2008). For instance, the Tutak Gneiss dome within the HP/LT belt is cored by gneisses and granites for which  $^{40}\text{Ar}/^{39}\text{Ar}$  dating yielded ages of 180 Ma and 77 Ma (Sarkarinejad and Alizadeh, 2009). In the Cheh–Galatoun (Quri) metamorphic mélange, few tens of kilometers to the east of the Neyriz obducted complex, amphibolites, garnet-bearing amphibolites and some eclogites or kyanite schists are exposed (Sarkarinejad et al., 2009).  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of the Quri amphibolites yields ages of ~91 Ma and 112–119 Ma in biotite gneiss. The good agreement found between the cooling ages in the Neyriz Ophiolitic complex and the HP tectonic mélange in the SSZ suggest a coeval episode of exhumation (Khadi

vi et al., 2012) during changes in subduction boundary conditions as inferred for the exhumation of blueschists in the southern Zagros (Agard et al., 2006). The tectonic position of the HP rocks with respect to the Neyriz ophiolites is debated. Indeed, it is still unclear whether the metamorphic mélange of SSZ is part of the upper Eurasian plate or the lower Arabian plate, mostly because of the obliteration of original structural relationships by subsequent deformation events. The HT/LP belt to the north (Fig. 2) is presumably older and probably results from arc-related regional metamorphism and magmatism (Sarkarinejad and Azizi, 2008). The metamorphic rocks are unconformably overlain by the Lower Cretaceous Orbitolina limestones, typical of the Central Iran sedimentation (Stocklin, 1974). The Late Cretaceous tectonic mélange of Neyriz is likely in a structural position equivalent to the Eocene magmatic assemblage in Kermanshah (ETMD, Agard et al., 2005, 2011). In NW Zagros, magmatism renewed in the Eocene as shown by gabbroic intrusions dated at 34 Ma (Leterrier, 1985) and by granitic intrusions (Gaiduh granite) occurred (Rachidnejad-Omran et al., 2002). A significant number of middle-upper Eocene apatite fission-track cooling ages were also determined from the Sanandaj–Sirjan zone (Homke et al., 2010) and could be correlated with the same Eocene fission-track ages reported from the Zagros foreland sedimentary rocks (Khadiji et al., 2012). The Miocene emplacement of the Sanandaj–Sirjan units along the MZT is revealed by thrusting of the Cretaceous limestones over Eocene and Miocene sedimentary rocks (Houshmand Zadeh et al., 1990). In addition, the Neogene clastics and red beds of the SSZ unconformably overlying metamorphic and magmatic rocks are folded, indicating that the emplacement of the SSZ was accompanied by deformation in the upper plate (Fig. 3).

### 2.4. The Urumieh–Dokhtar Magmatic Arc (UDMA)

The Urumieh–Dokhtar Magmatic Arc (UDMA; Fig. 1) located between the SSZ and Central Iran is striking parallel to the Zagros Mountains. The oldest rocks in the UDMA are calc-alkaline magmatic rocks, which cut across Upper Jurassic formations and are overlain unconformably by Lower Cretaceous fossiliferous limestones. The UDMA is characterized by Tertiary magmatism, showing the migration of magmatic activity from the SE (SSZ). Magmatism has been mainly active in the Eocene (Fig. 3) in association with the subduction of the Neo-Tethyan slab and continued for the rest of that period with a magmatic flare-up that lasted from 55 to 37 Ma (Verdel et al., 2011), including the Tarom plutonic belt of Berberian et al. (1982). The UDMA is composed of voluminous tholeiitic, calc-alkaline, and K-rich magmatic rocks. Paleogene volcanics and sedimentary rocks reach 3–8 km in the UDMA and the Alborz, indicating subsidence associated with back-arc extension (Ballato et al., 2011; Morley et al., 2009; Verdel et al., 2007, 2011; Vincent et al., 2005) and formation of Eocene core-complexes in central Iran (Moritz et al., 2006; Verdel et al., 2007). The flare-up ceased at ~37 Ma age, as indicated by zircon U/Pb and Ar/Ar ages on a gabbro intruding the Karaj Formation with a typical continental arc signature. In the Oligocene, magmatism changed to OIB-like volcanism that reflects asthenospheric-derived melting (Verdel et al., 2011). The youngest rocks in the UDMA consist of lava flows and pyroclastics that belong to Pliocene and Quaternary volcanic cones of alkaline nature (Berberian and Berberian, 1981). Post-collisional Pliocene–Quaternary volcanism was suggested to result from lithosphere delamination beneath the overthickened Iranian Plateau (Hatzfeld and Molnar, 2010) or breakoff of the Neoethyan slab (Jahangiri, 2007; Omrani et al., 2008; van Hunen and Allen, 2011).

## 3. Re-appraisal of the lithosphere-scale collision structure of the Zagros

In this section, we present updated geophysical constraints on the deep lithospheric structure and shallow crustal-scale features of the Zagros collision, which are essential to better understanding the



geodynamic evolution of the Arabia/Eurasia collision and mountain building in the Zagros.

### 3.1. Geophysical constraints on continuous/discontinuous Arabian slab

Receiver function analysis of teleseismic earthquakes performed along two 500 km-long profiles provide constraints on the crustal thickness in the Zagros belt and the adjacent Iranian Plateau (Paul et al., 2006; Paul et al., 2010) (Figs. 2 and 4). Beneath the ZFB and the High Zagros, the Arabian crust is  $42 \pm 2$  km thick and the thickness of the lower crust is  $\sim 25$  km (Hatzfeld et al., 2003). The consistency with the unthickened portion of the Arabian margin (Gök et al., 2008) indicates that the crust has not yet been significantly thickened in this part of the Zagros or has a wavelength too small to be fully compensated by the flexure of the continental margin. The average geothermal crustal gradient in the Arabian crust has been constrained to  $15\text{--}24$  °C/km in agreement with thermochronometric studies (Gavillot et al., 2010; Homke et al., 2010; Khadivi et al., 2012) and tectonic modeling (Mouthereau et al., 2006). It is worth noting that such a low geothermal gradient is compatible with the presence of earthquakes in the lower crust, down to depths of 30 km (Roustaei et al., 2010; Talebian and Jackson, 2004; Tatar et al., 2004).

Beneath the Sanandaj–Sirjan metamorphic belt the Moho deepens to a depth comprised between  $69 \pm 2$  km (central Zagros) and  $56 \pm 2$  km (northern Zagros) (Paul et al., 2010). Northwards, beneath Central Iran the crustal thickness reduces to  $\sim 50$  km. There, in contrast to the Arabian mantle, which displays shield-like P-wave velocities, lower seismic wave velocity suggest an anomalously thin upper mantle lithosphere, warmer than the Arabian one at least down to depths of 100 km (Chang et al., 2010; Kaviani et al., 2007). Mantle thinning has been interpreted to be possibly related to the removal of the mantle lithosphere following the thickening of the Iranian continent (Hatzfeld and Molnar, 2010; Maggi and Priestley, 2005). Such finding has however been challenged by more accurate velocity estimates showing no evidence for a high-velocity anomaly zone below 100 km and therefore do not seem to support the convective removal of the mantle lithosphere beneath the plateau (Kaviani et al., 2007; Paul et al., 2010). Receiver function tomography further reveals the occurrence of a strong reflector with a low-dip to the NE that is in continuation with the Main Zagros Thrust at surface, penetrating some 250 km below Central Iran. Paul et al. (2010) interpreted this reflector as the plate boundary between Eurasia and Arabia, revealing the underthrusting of the Arabian plate. This finding is in agreement with null shear wave splitting beneath the Zagros and the Sanandaj–Sirjan suggesting that both regions are underlain by the same Arabian lithosphere (Kaviani et al., 2009).

Global tomography data presented in recent studies (Agard et al., 2011; Vergés et al., 2011b) show instead a sharp boundary below the Arabia and Eurasia plate boundary in NW Zagros, dipping about  $50^\circ$  to the NE. Together with other recent tomographic constraints (Chang et al., 2010), these tomographic models reveal a discontinuous Neo-Tethyan slab beneath the NW Zagros (west of  $51^\circ\text{E}$ ), in agreement with previous tomography data and inferred slab-breakoff (e.g., Hafkenscheid et al., 2006). In contrast, the Arabian lithosphere, as imaged by recent improved tomographic models, is continuous in the central Zagros (Chang et al., 2010). Simmons et al. (2011), using a novel multi-event location approach and 3D-ray tracing (Myers et al., 2011), image a continuous fast-velocity domain in the shallow upper mantle beneath Iran (Fig. 4). The fast-velocity anomaly is particularly well-defined in southern Zagros (Fars) and can be followed continuously northwards over 500 km, indicating the underthrusting of the Arabia lithosphere beneath Central Iran over a distance compatible with the Miocene age of the collision. Whether the Arabian lithosphere is continuous or not in the narrow region comprised between the NW Zagros (west of  $51^\circ\text{E}$ ) and the central Zagros (east of

$52^\circ\text{E}$ ), is however unclear from the available constraints. A high-velocity anomaly is identified beneath central Iran (Keshvari et al., 2011; Shomali et al., 2011) and could indicate a continuous slab.

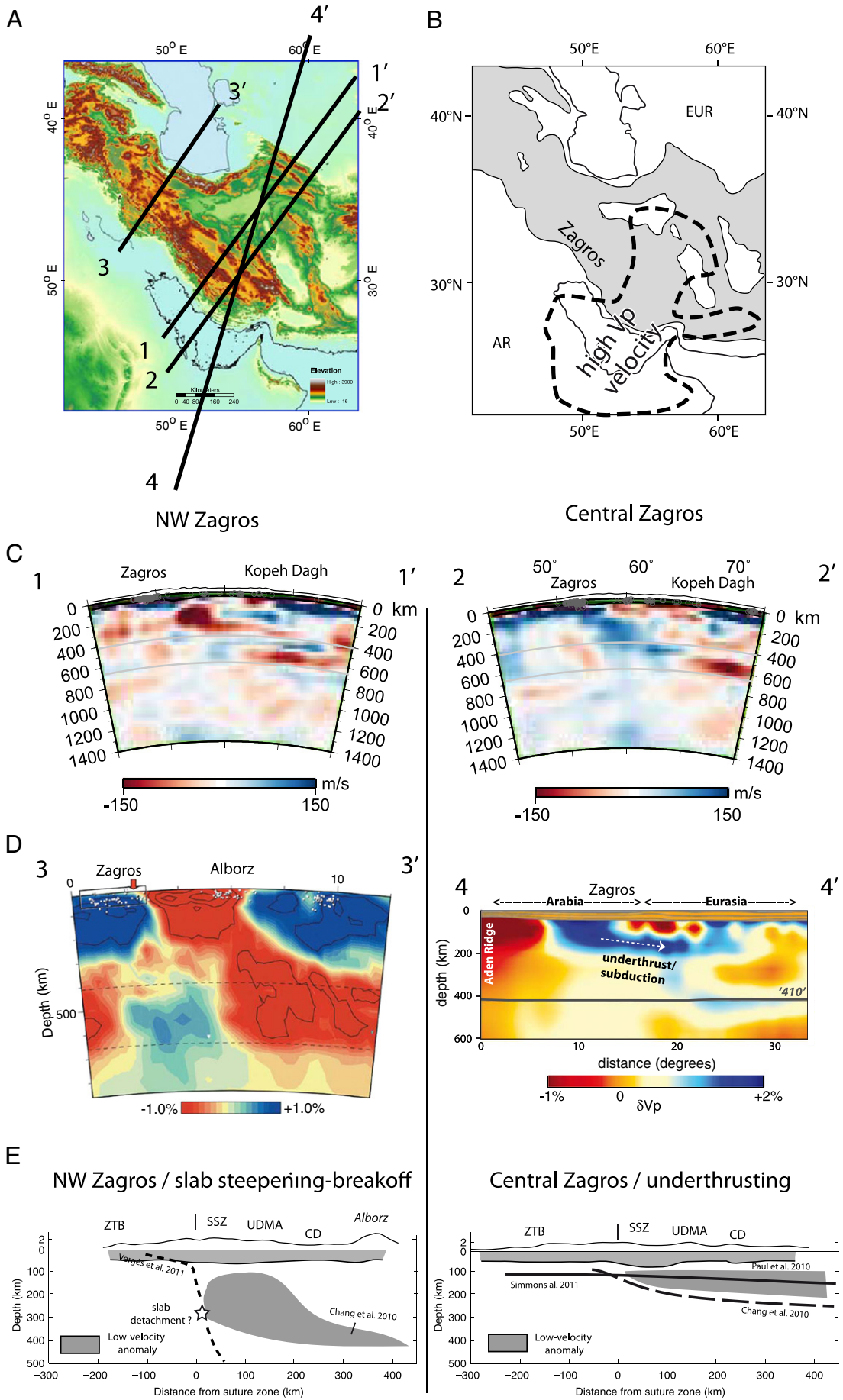
Based on these tomography data it can be proposed that slab steepening or slab breakoff occurred in NW Zagros. This would be in agreement with the breakoff stage inferred in Anatolia, which possibly led to the formation of the North Anatolian Fault at  $\sim 12$  Ma (Authemayou et al., 2006; Faccenna et al., 2006; Regard et al., 2005). To the south, however, tomographic models with improved resolutions (Chang et al., 2010; Paul et al., 2010; Simmons et al., 2011) show a cold slab beneath central Iran. It is characterized by low dip angles of the underthrusting/subducting Arabian lithosphere (Fig. 4). These data do not support a recent slab breakoff in the central Zagros, as it was anticipated from the occurrence of upper Miocene–Pliocene–Quaternary adakites in the UDMA region (Jahangiri, 2007; Omrani et al., 2008). At best, tomographic constraints could support an older Neo-tethyan slab detachment at  $\sim 30$  Ma or older (e.g., Hafkenscheid et al., 2006).

### 3.2. Zagros structure: cover folding and basement-involved shortening

Folding of the sedimentary cover in the Zagros was permitted by the basal Hormuz evaporitic layer present in the Fars arc or by an equivalent layer in the Dezful embayment and Lorestan arc (Fig. 2). This thick (1–2 km) and weak salt-bearing formation has led us to interpret the Zagros Folded Belt as a thin-skinned fold-thrust belt (Davis and Engelder, 1985). This major décollement is obviously at the origin of the periodic folds that characterized the physiography of the ZFB. However, large-scale uplifts in the basement were also early described as major “geo-flexures” by Falcon (1961). For instance, a structural relief of 3 km is observed across the Mountain Front Flexure in SE Lorestan and NW Dezful Embayment (Emami et al., 2010; Falcon, 1961; Sherkati et al., 2006) reaching almost 6 km in the central Izeh Zone. In the Fars arc the basement uplifts are characterized by several topographic steps displaying larger wavelengths (40–100 km) that are associated with stepwise structural elevation of synclinal bottoms (Leturmy et al., 2010; Mouthereau et al., 2006). These steps are geographically associated with the inferred positions of well-known active basement thrusts such as the Mountain Front Fault or Surmeh Fault, along which major historical and instrumental earthquakes have been reported (Berberian, 1995). These morpho-structural features support basement faulting in the Zagros. Similar tectonic relationships were reported throughout the ZFB including the Izeh zone (Sherkati and Letouzey, 2004; Sherkati et al., 2006), the Lorestan (Blanc et al., 2003) and the south-eastern Fars (Molinari et al., 2005a).

In one way or another, cross-sections crossing the Zagros incorporate both thin-skinned and thick-skinned structures to interpret the observations (Fig. 2), although an alternative thin-skinned structural interpretation has been proposed for the geometry and kinematics of the ZFB (McQuarrie, 2004). In this interpretation the structural relief associated with the Mountain Front Fault is attributed to the accumulation of the Hormuz salt e.g., beneath the Pusht-e Kuh arc front. However, it has been shown that structural models making the Hormuz evaporitic layer the main basal décollement is unlikely to maintain the regional topography (Mouthereau et al., 2006).

Attempts to accurately determine the centroid depths of waveform-modeled earthquakes locate the main events in the upper part of the crust (Talebian and Jackson, 2004; Tatar et al., 2004). A noticeable advance on this question has been brought by the reappraisal of the seismicity distribution in the SE Zagros associated with the 2006 sequence of Fin earthquakes (Mw 5.7, 5.5, 5.2, 5.0, 4.9) (Roustaei et al., 2010) and the two events (Mw 5.8, 5.9) that occurred in the Queshm island (Nissen et al., 2010). These studies reveal that the main rupture occurred in the competent cover with reverse slip restricted between about 4 and 10 km but



failed to propagate across the Hormuz salt, producing a cluster of aftershocks in the crystalline basement (10–30 km; Roustaei et al., 2010). A more regional re-appraisal suggests that the basement is less seismogenically active and may deform by aseismic creep (Nissen et al., 2011). This may reveal that the basement is rheologically stronger than the cover and therefore does not contradict the long-term involvement of the basement in deformation. Seismogenic basement deformation in the central Zagros appears localized across the Mountain Front Fault and the Surmeh Fault (Fig. 2). This localized, active, deformation in the upper brittle crust may root at depth into mostly aseismic ductile shear zones at lower crustal levels (Fig. 2).

The above constraints support the proposed structural cross-sections of the Zagros in which the long-term shortening is achieved by superimposed cover folding above the décollement in the Hormuz salt that is cut occasionally by active basement thrusts (Blanc et al., 2003; Emami et al., 2010; Molinaro et al., 2005a; Mouthereau et al., 2007b; Sherkaty and Letouzey, 2004; Sherkaty et al., 2006).

#### 4. Cenozoic deformation history of the Zagros collision

##### 4.1. Temporal constraints on the onset of Arabia–Eurasia collision

The age for initiation of the Arabia–Eurasia collision has been constrained between ~64 Ma (Berberian and King, 1981), using the end of ophiolite obduction, and ~5 Ma using the angular unconformity between Bakhtyari conglomerates and the underlying Agha-Jari Formation (Falcon, 1974a,b). None of these approaches provides a date for the first time Arabian and Eurasian continental crusts came into contact in response to convergence.

Arabia–Eurasia paleotectonic maps indicate that collision should occur during Miocene times (Barrier and Vrielynck, 2008; McQuarrie et al., 2003) in agreement with the progressive closure of the Mediterranean–Indian water gate (Reuter et al., 2009). This timing is consistent with the geological evidence of the final closure of the Neo-Tethyan ocean between Arabia and Eurasia at ~20 Ma in the Bitlis suture, to the NW of the Zagros (Okay et al., 2010). In the High Zagros, the emplacement of the gabbroic Tah intrusion (Rb–Sr age of ~34 Ma) of the Gaveh-Rud domain (upper plate, southern SSZ), before the deposition of the ~20–18 Ma Asmari Formation provided a consistent age for the onset of the collision (Fig. 3; Agard et al., 2005; Braud, 1987; Letierrier, 1985).

One large geodynamic event in the Zagros basin was recorded by the initiation of foreland basin subsidence caused by the Arabian plate deflection. The deposition of Shahbazan carbonates at ~34 Ma (Homke et al., 2009) in the NW Zagros outlines the onset of subsidence in the Zagros basin, which may be related to the preliminary stage of the plate flexure (Fig. 3). This event is not recorded in the Fars region likely because only the proximal part of the margin is preserved (Mouthereau et al., 2007b). As depicted in Fig. 3, the deposition of the shallow-marine Asmari carbonates, unconformably overlying the older Shahbazan or Jahrom Formation to the north (Homke et al., 2009) and the younging of the Asmari Formation forelandward (James and Wynd, 1965) argue for the migration of basin depocenters with the propagation of thrust loading (Mouthereau et al., 2007b). The reappraisal of the sedimentology and stratigraphy of the Asmari Formation reached the same conclusion; the Asmari Formation is syn-collisional and deposited in a flexural basin (van Buchem et al., 2010). The Asmari Formation has been dated to the middle-early Miocene (~20–18 Ma) using  $^{87}\text{Sr}/^{86}\text{Sr}$  isotopic dating in the Dezful embayment (Ehrenberg et al., 2007) and in the Lorestan

arc (Saura et al., 2011), which therefore provides a date for the onset of the collision. This interpretation is consistent with the onset of coarse-grained deposition in foreland succession at ~23 Ma in the High Zagros (Fakhari et al., 2008), at ~15 Ma in the ZFB (Khadivi et al., 2010) and at 13.5 Ma in the SE Lorestan region (Emami, 2008). The early Miocene coincides with the increase in sediment clastics influx from the growing orogen, e.g., the Razak Formation dated at 19.7 Ma (Khadivi et al., 2010). A detrital AFT cooling age of 25 Ma, reported in the 19.7 Ma Razak Fm (Khadivi et al., 2012) of the central Zagros, points to the onset of exhumation in the early Miocene (Fig. 3). This could have resulted from thrust motion during the plate suturing in the Zagros between 34 Ma and 20 Ma (Agard et al., 2005; Braud, 1987).

Petrographic constraints indicate that magmatic rocks behind the UDMA recorded the transition from post-collisional calc-alkaline to alkaline volcanism, between 30 and 23 Ma, according to new zircon U/Pb dates (Aghazadeh et al., 2011). In the Alborz, Guest et al. (2007) documented basalts, in the Qom Formation, with Ar/Ar wholerock age of ~33 Ma. Geochemistry of these basalts shows an OIB trace element signature whereas a gabbro dated at ~37 Ma (Ar/Ar on biotite) has an arc signature (Verdel et al., 2011). The demise of arc-related magmatism in Central Iran, during the late Eocene–early Oligocene, was taken to indicate the timing of the collision with Arabia (e.g., Allen and Armstrong, 2008; Ballato et al., 2011; Vincent et al., 2005). Low-temperature thermochronology and provenance studies in the Western Great Caucasus have consistently pointed to an early Oligocene age for the final closure of the Neotethys (Vincent et al., 2007). The 34–23 Ma interval is contemporaneous with the onset of inversion of Paleocene–Eocene backarc extensional basins of central Iran (Ballato et al., 2011; Morley et al., 2009). Taken together these data confirm that the active subduction of the Arabian slab ceased in the early Oligocene contemporaneous with the initiation of SW slab retreat and the transition to collision in the Zagros.

In summary, constraints on the timing of Neo-Tethyan ocean consumption, Zagros sediment provenance and stratigraphy as well as the age of arc-related magmatism in the Iranian microplate support initiation of the collision at 34 Ma with the final closure of the Neo-Tethys at 20 Ma (Fig. 3). Such a long transitional period is interpreted to reflect the switch from underthrusting of the stretched Arabia continental margin to the initiation of crustal thickening of the unstretched portion of the Arabian lithosphere (Ballato et al., 2011; Mouthereau, 2011).

##### 4.2. Timing of folding: dating of growth strata in the Zagros folded belt

The timing of folding in the Zagros has been originally associated with the coarse-grained deposits of the Bakhtyari Formation unconformably overlying Agha Jari Formation and older sedimentary units (Falcon, 1974a,b; James and Wynd, 1965; Stocklin, 1968). This led to assign a late Pliocene or younger age for the onset of deformation. Field evidence for more complex relationships between the conglomeratic deposition and folding (Fakhari et al., 1977; Fakhari et al., 2008; Khadivi et al., 2010; Mouthereau et al., 2007a,b) argued for the need to gain more accurate stratigraphic constraints.

Accurate dating of deformation in fold-thrust belts requires the preservation of coeval tectonic/stratigraphic relationships such as growth strata geometries (Fig. 5). While such geometries can be inferred in the Zagros either in the field or on seismic lines only few good field sections have been dated by magnetostratigraphy (Homke et al., 2004; Khadivi et al., 2010). Homke et al. (2004) were

**Fig. 4.** Tomographic results in the Middle East shown as vertical cross-section maps in the NW Zagros and central Zagros. A) Location of cross-sections. B) Distribution of a high velocity zone depicting underthrust Arabian lithosphere after Simmons et al. (2011). C) Cross-sections in NW and central Zagros obtained by joint inversion tomography after Chang et al. (2010). D) –Cross-section across the NW Zagros obtained by joint inversion tomography after Vergés et al. (2011a,b). E) Cross-section in central Zagros showing the underthrust Arabian slab, obtained after Simmons et al. (2011). F) Interpretation sketch based above tomography data and receiver function for the crustal thickness (Paul et al., 2006, 2010). They consistently show a steep slab in NW Zagros whereas a flat subduction in central Zagros is depicted, thus indicating the underthrusting of the Arabian margin. Abbreviations: ZFB (Zagros Folded Belt); SSZ (Sanandaj–Sirjan Zone); UDMA (Urumieh Dokhtar Magmatic Arc); and CD (Central Domain).

the first to conduct magnetostratigraphic studies of pre-growth strata (few tens of meters of Gashsaran Fm. and 800 m of fluvial Agha Jari Fm.) and growth strata (1200 m of fluvial Agha Jari Fm and Bakhtyari alluvial Fm.) at the front of the Pusht-e Kuh arc (Anaran anticline), along the Changuleh growth syncline, an area positioned between 200 m and 500 m above sea level (Fig. 2). This work provided constraints on the timing of folding at 7.6 Ma and Mountain Front Fault (MFF) uplift at 2.5 Ma (Fig. 3). In order to yield constraints closer to the High Zagros, a second section was dated by magnetostratigraphy in the northern flank of the Chahar–Makan syncline at an altitude of ~2500 m, 20 km to the NW of Shiraz, in the North of the Fars arc and just south of the inferred position of the High Zagros Fault (Fig. 5). The pre-growth is characterized by 500 m-thick Razak coastal sabkha deposits dated at 19.7 Ma and 400 m-thick Agha Jari deltaic sandstones dated at 16.6 Ma. The growth patterns in the alluvial conglomerates of Bakhtyari 1 reveal a minimum age for the initiation of folding of 14.8 Ma (Khadivi et al., 2010). Interestingly, the Bakhtyari 1 (Bk1) of the Chahar–Makan syncline has subsequently been folded during the development of the Derak anticline in the south. The folding event is outlined by one angular unconformity between Bk1 and the slightly N-dipping conglomeratic layers of the Bakhtyari 2 Formation (Fig. 5). The top of the folded Bakhtyari 1 conglomerates can be dated at ~12.4 Ma by extrapolating accumulation rates. This second event of folding has been tentatively correlated from fold to fold across the entire Fars area, thus arguing for a rapid phase of fold amplification at 10–5 Ma (Mouthereau et al., 2007b), but this obviously

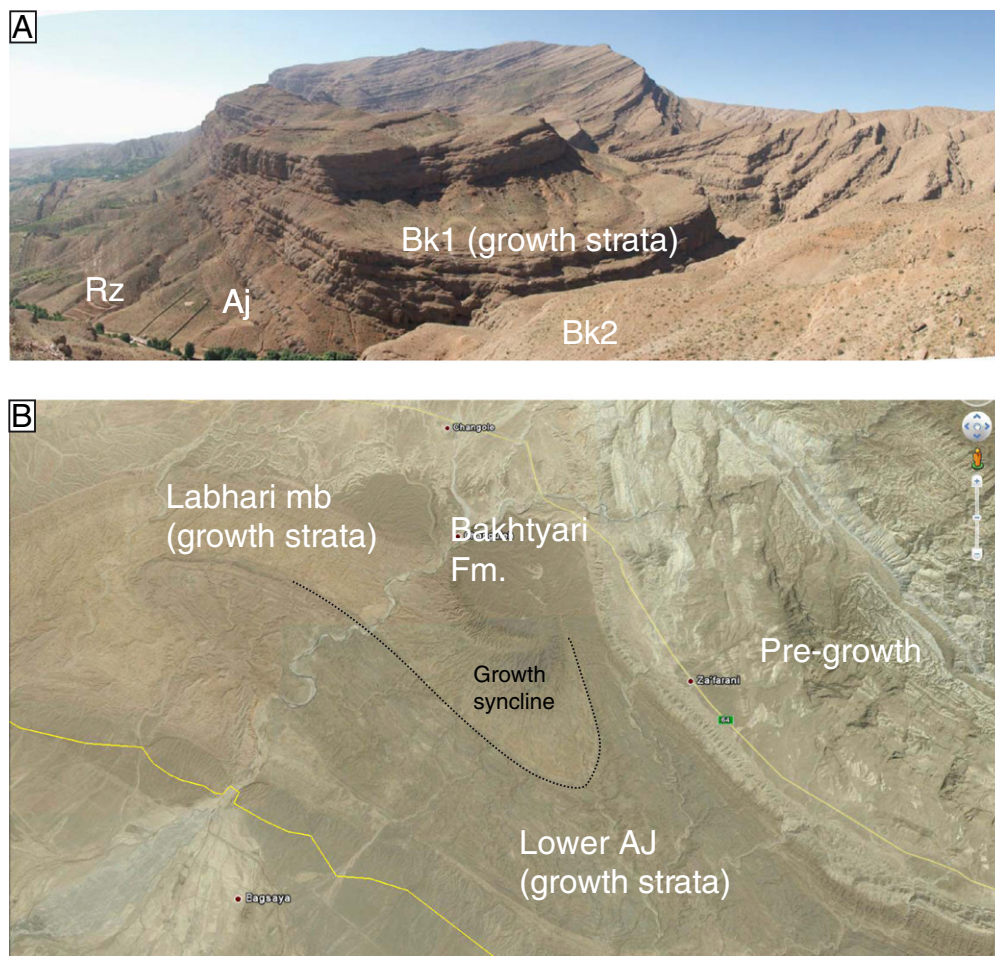
needs to be confirmed by more direct datings. A phase of fold tightening at ~2–3 Ma has also been inferred based on the approximate age of the youngest flat-lying Bakhtyari conglomerates (Mouthereau et al., 2007b), which are lying unconformably in most parts of the Zagros and fit with the original definition of the Bakhtyari Formation (Falcon, 1974a,b; James and Wynd, 1965; Stocklin, 1968).

Magnetostratigraphic dating in the Chaman–Goli syncline, in the hangingwall of the HZF (Izeh Zone), yielded a slightly younger age of 11 Ma for possible growth units at the Agha Jari/Bakhtyari transition (Emami, 2008). Biostratigraphic dating of growth strata in the Amiran basin in NW Zagros (Lorestan) yielded a Maastrichtian age for the oldest folding event (Saura et al., 2011) in relation to obduction.

It is apparent from tectonic/stratigraphic relationships that the main phase of folding in the ZFB occurred after ca. 12 Ma. Since then, the tightening of folds and the interactions with basement faulting likely resulted in a complex sequence of folding, including out-of-sequence deformation.

#### 4.3. Relative timing between cover and basement deformation in the ZFB

The relative timing between cover folding and basement shortening in the ZFB is worth to be studied since it reflects the evolution of collisional stress build-up and provides constraints on the sequence of deformation and topographic evolution. The current debate focuses on whether shortening in the basement occurred first and was later



**Fig. 5.** Examples of growth strata in the northern ZFB and at the mountain front. A) Growth strata geometry on the northern flank of the Derak anticline (northern ZFB, Fars region). The base of the growth strata has been dated to ~15 Ma by magnetostratigraphy (Khadivi et al., 2010). B) Satellite image of the Changuleh growth syncline (GoogleEarth©) at the southern front of the Zagros belt. The base of these growth strata has been dated to 8.1–7.2 Ma by magnetostratigraphy (Homke et al., 2004).

transmitted to the cover, or occurred afterwards. Alternatively, both cover and basement deformation may occur contemporaneously.

Local variations of subsidence, lithofacies, and disconformities in Miocene foreland deposits outline the segmentation of foreland subsidence patterns in the ZFB. In the Fars, the inhomogeneous pattern of Miocene sediment thickness (Fars Group) coincides with the occurrence of NW-trending depocenters that are related to basement block faulting. Together with field evidence for the pinching out of sedimentary markers (Guri limestones) in the vicinity of an inferred basement strike-slip fault (Mouthereau et al., 2007a,b), these constraints support tectonic inversion of the rifted continental basement at 16–11 Ma (Mouthereau et al., 2006). The analysis of the early fracture pattern and basin architecture at the time of or immediately following deposition of the Asmari Formation (Ahmadhadi et al., 2007; Ahmadhadi et al., 2008) also supports the occurrence of early compressional reactivation of basement faults within the Zagros before significant involvement of the cover.

Such an interpretation might not be valid along the strike of the ZFB especially in regions where salt diapirism has been particularly active such like the SE Fars area (e.g., Jahani et al., 2009). In the High Zagros, the early development of basement faulting during the Paleocene–early Eocene is indicated by the 1.3-km basement uplift, which controlled the geometry of deposition of the Amiran–Kashkan sequences (Homke et al., 2009) (Fig. 3). On the other hand, there are numerous structural arguments that active basement faulting cut across the original fold geometry, indicating that basement-involved shortening also postdates cover folding (Casciello et al., 2009; Molinaro et al., 2005a,b; Mouthereau et al., 2007b). Stratigraphic constraints available at the front of the Pusht-e Kuh arc place the initiation of basement faults along the Mountain Front Flexure at 5.5 Ma, thus postdating the initiation of cover folding in this region (Emami et al., 2010). The same timing is proposed in the SE Fars (Molinaro et al., 2005a). The late stage of basement strike-slip faulting that resulted in the offset of cover fold axes in the Central ZFB likely initiated during the Pliocene (2–5 Ma), as inferred from Quaternary slip rates along the Kazerun Fault (Authemayou et al., 2006, 2009).

The current seismicity and GPS data provide evidence for both deep-seated basement and cover are actively deforming. Based on evidence that basement faulting occurred before and after folding, or synchronously as inferred from present-day deformation and faulting, we speculate the ZFB evolved as a superimposed thin-skinned and thick-skinned fold-thrust belt. As in many fold belts characterized by weak basement–cover interface, the deformation in the basement occurs in the hinterland and is transferred towards the foreland in the cover system (see restored section for Lorestan in Vergés et al., 2011b). The resulting cover shortening was therefore achieved through a combination of detachment folds and forced folds (i.e., folds forced by slip along underlying basement faults) (Abdollahie et al., 2006; Blanc et al., 2003; Mouthereau et al., 2007b; Oveisi et al., 2009; Ramsey et al., 2008; Sattarzadeh et al., 2000). The sequence between the superimposed structural levels was probably dependent on many controlling factors such as the tectonic inheritance and the feedback with sedimentation/erosion pattern.

#### 4.4. Timing of folding and thrusting in the High Zagros

Here, we focus on the timing of deformation in the High Zagros, structurally below the MZT and the plate suture. In this part of the Zagros, the timing of shortening is not well constrained mainly due to the lack of syntectonic Cenozoic series. The presence of Oligo-Miocene limestones unconformably overlying folded Mesozoic strata reveals that folding, uplift and erosion occurred before deformation developed in the ZFB, as early as the Eocene (Khadivi et al., 2012), following a classical forward-propagating sequence as noticed by Hessami et al. (2001). As pointed out earlier, this is in agreement

with evidence of Paleocene–early Eocene deformation sealed by Kashkan Formation (Homke et al., 2009) in NW Zagros (Fig. 3). The coarse-grained Shalamzar Bakhtyari Formation with intercalated fossiliferous marine deposits unconformably overlies the folded Razak, Agha Jari and non-marine Bakhtyari formations in the hangingwall of the MZT (Figs. 2 and 3). The biostratigraphic dating yields an early Miocene age (Aquitainian ~20–23 Ma) for the onset of folding (Fakhari et al., 2008).

Apatite (U–Th)/He (AHe) thermochronology data across the Lajin and the Dinar thrusts of the High Zagros indicates that thrusting occurred between 19 and 15 Ma and 12 and 8 Ma, respectively (Gavillot et al., 2010). Similar AHe ages of 19–15 Ma were recovered from the Bakhtyari conglomerates dated by Fakhari et al. (2008) in the footwall of the MZT. Because these conglomerates are essentially made with clasts originated from the hangingwall of the MZT there is no doubt that the fault was active by early Miocene times at ~20–23 Ma. The early Miocene (Oligocene?) emplacement of the MZT is supported by the overthrusting of the Cretaceous limestones over Eocene and Miocene sediments of the High Zagros (Khadivi et al., 2012). It is also confirmed by the structural position of the SSZ units on top of Miocene flysch units in the High Zagros near Kermanshah (Agard et al., 2005).

To summarize, deformation propagated as a typical outward sequence from the HZ in the Eocene to the inner ZFB in the middle Miocene (20–8 Ma). In contrast, within the ZFB, folding/thrusting propagated rapidly throughout the width of the range at 12–5 Ma, leading to an apparent synchronicity between cover and basement deformation, although the current shortening is mainly absorbed across the most frontal folds (Oveisi et al., 2009; Walpersdorf et al., 2006).

## 5. Cenozoic uplift and exhumation history of the Zagros collision

### 5.1. Syn-collisional and pre-orogenic cooling and exhumational events in the Zagros: low-temperature thermochronologic constraints

A few thermochronometric data have provided insights on the cooling history in the Zagros (Fig. 3). Detrital apatite fission-track thermochronology carried out on Miocene foreland sedimentary rocks of the Zagros Folded Belt documents a regional cooling event ranging between 25 Ma in the northern Fars region (Razak Fm deposited at 19.7 Ma in Chahar-Makan syncline; Khadivi et al., 2012) and 22 Ma in the Lorestan (Lower Agha Jari Fm dated at 12.8 Ma in the Zarrinabad syncline; Homke et al., 2010). Taking into account a closure temperature of 110 °C and assuming a geotherm of 15–24 °C/km (Gavillot et al., 2010; Homke et al., 2010; Mouthereau et al., 2006), one estimates that roughly 4.5–7 km were exhumed since the early Miocene. Based on pre-collisional zircon (U–Th)/He ages (Tc ~ 180 °C) a consistent maximum exhumation of 7–9 km in the High Zagros is inferred (Gavillot et al., 2010).

A petrographic and provenance analysis of early-middle Miocene rocks suggests that such a Miocene cooling event most likely resulted from uplift and erosion in the hangingwall of the MZT. This is supported by an AFT cooling age of 27 Ma reported from a gneiss sample of the Dorud metamorphic complex of the SSZ (Homke et al., 2010). Moreover, petrographic studies in sediments deposited at this time support the erosion of the radiolaritic and ultramafic complex, thus indicating deformation in the High Zagros and more generally the exhumation of the suture zone near the Main Zagros Thrust. Such a thrust sequence is consistent with apatite (U–Th)/He ages-elevation profile in the Lajin thrust, which provided evidence for a rapid cooling event between 19 Ma and 15 Ma (Gavillot et al., 2010) in the High Zagros. This result matches the inferred exhumation in southern SSZ and folding in the northern ZFB. In the Crush Zone (Kermanshah), unpublished AHe data suggest burial between 16 and 12 Ma and

exhumation by deep-seated imbricate thrusts (underplating) at 12 Ma (Wrobel-Daveau et al., 2011).

Among AFT grain ages recovered from the foreland sedimentary rocks, both Jurassic to Early Cretaceous cooling ages are well defined across the foreland regions (Homke et al., 2010; Khadivi et al., 2012). These ages reveal the long-term magmatic evolution during the early subduction of the Neo-Tethyan slab beneath the SSZ. Paleocene–Eocene cooling ages are consistent with the well-defined episode of extensional magmatic flare-up in Central Iran, which lasted from 55 Ma to 37 Ma (Berberian and Berberian, 1981; Berberian and King, 1981; Verdel et al., 2011).

The preservation of Mesozoic, Eocene or early Miocene cooling ages allows a rough estimate of the maximum exhumation in the ZFB. Based on the reconstructed total thickness of the synorogenic Miocene sediments, a maximum exhumation of 2.5 km can be inferred (Khadivi et al., 2012), yielding exhumation rates of 0.2 km/Myr if an age of 12.4 Ma is assumed for the onset of regional-scale folding.

To summarize, detrital thermochronologic analyses in the Miocene–Pliocene sedimentary rocks of the Zagros foreland show a main syn-collisional exhumational event in the hinterland at ~25 Ma in the SSZ (Fig. 3). SSZ units, characterized by Mesozoic and Paleocene–Eocene apatite fission-track cooling ages, therefore came into contact with the Arabian margin as soon as 19.7 Ma, thus providing constraints on the age of the collision. At ~12 Ma, exhumation that resulted from the southward propagation of deformation in the ZFB occurred in agreement with reset AHe ages found in the Crush Zone (Wrobel-Daveau et al., 2011). This date would provide additional constraints on the time at which the Arabian margin stopped underthrusting below central Iran and started to thicken, building the Zagros orogenic wedge in front of the Iranian plateau.

### 5.2. Temporal constraints on uplift of the Zagros belt and some implications

The timing of Zagros uplift is primarily constrained by the age of the youngest marine sediments dated in Iran. In this respect, the gradual transition from marine sedimentation to non-marine coarse-grained deposits in nearshore fan deltas reveals that until ~15 Ma the northern Zagros Folded Belt was close to sea-level (Khadivi et al., 2010). This is consistent with the age of the last marine limestones in the Zagros and the Iranian Plateau dated to the Burdigalian (15.9–20.4 Ma) (Harzhauser et al., 2007; Schuster and Wielandt, 1999). The unconformity between folded Bakhtyari 1 conglomerates and the flat-lying Bakhtyari 2 conglomerates in the northern ZFB (Fars) is roughly dated to 12.4 Ma by extrapolating accumulation rates from magnetostratigraphic dating (Fig. 5). This angular unconformity outlines a tectonic episode that could indicate an episode of uplift after which the Zagros rose to a maximum altitude of ~2 km.

AFT cooling ages ( $T_c \sim 110^\circ\text{C}$ ) obtained from detrital fission-track analysis in foreland sediments ranging between 19.7 Ma (Khadivi et al., 2010) and 3 Ma (Homke et al., 2010) show a lack of grains younger than ~20 Ma. Abundant older pre-collisional Eocene and to less extent Mesozoic grain-age populations instead reveal a moderate exhumation. Only thermochronometers with lower closure temperature like (U–Th)/He on apatites ( $T_c \sim 70^\circ\text{C}$ ) are able to detect cooling events during the Late Miocene (12–8 Ma) in association with the High Zagros Fault (e.g., Dinar thrust; Gavillot et al., 2010) and exhumational AHe age pattern in the NW Zagros after 12 Ma (Wrobel-Daveau et al., 2011).

Clay mineralogy and sedimentology of the synorogenic deposits both emphasize that climate was chiefly warm and dry with no marked changes between 19.7 Ma and 13.8 Ma in the northern Zagros (Khadivi et al., 2012). The widening of the Zagros–Iranian Plateau through the development of new thrusts or folds at 12.4 Ma,

in the Zagros, was likely favored by the imbalance between the limited erosion under the prevailing dry conditions and tectonics. The development of internally drained areas in the northern Zagros–Iranian Plateau region (Khadivi et al., 2012; Mouthereau et al., 2007; Walker et al., 2011) likely reduced the erosional capacity within channels by disconnecting streams from a stable regional base level (e.g., Sobel et al., 2003). This prevented the establishment of a positive feedback between tectonics and erosion and therefore facilitated Zagros uplift and plateau expansion. In this aspect, the Zagros may be viewed as an example of accretionary wedge that has not reached a topographic steady-state in the sense of Willett and Brandon (2002).

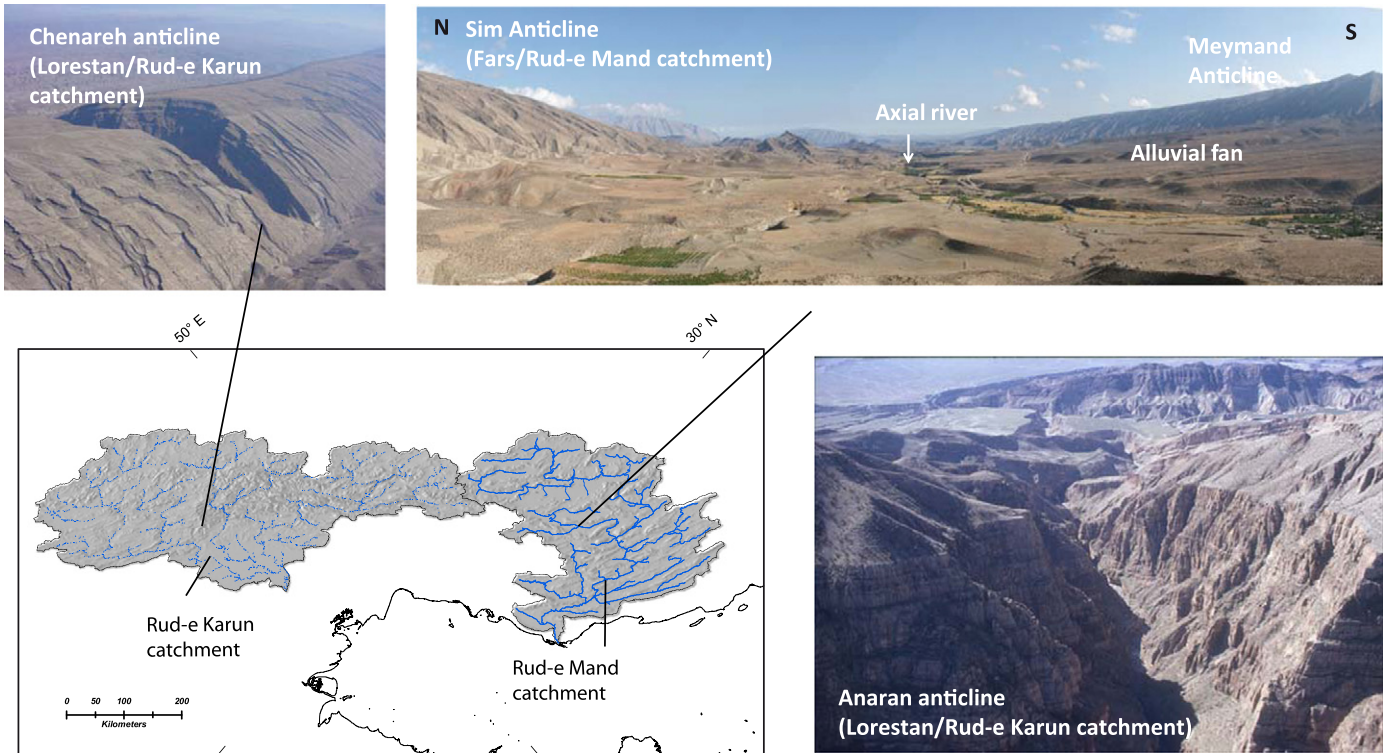
### 5.3. Fold development and evolution of drainage networks

To date a limited number of studies emphasized how fold morphology developed in the Zagros. Here, we discuss first-order constraints on the mechanisms and timing of fold-related drainage network development.

Drainage pattern in the Dezful area of the Zagros folded belt is characterized by spectacular river gorges cutting through Mesozoic and mid-Tertiary limestones along streams transverse to fold axis (Fig. 6). This pattern has originally been studied by Oberlander (1985) who proposed that such transverse streams resulted from the superposition of the planform geometry of continuous rivers that primarily developed in soft sediments and secondarily cut through exhuming resistant Mesozoic and mid-Tertiary limestones. By contrast, in the Fars region, rivers do not follow a simple south-flowing course across the folded belt, as transverse rivers have a course shorter than axial rivers (Fig. 6). Such a drainage pattern strongly suggests that rivers cannot keep pace with fold uplift as the stream power is not sufficient to incise the resistant Mesozoic limestones (Mouthereau et al., 2007b; Ramsey et al., 2008). The course of transverse rivers has been dammed and deviated by rapidly uplifted folds forcing rivers to flow axially. In the northern Fars, the river network appears to be also controlled by the differential uplift and tilting of the regional base level. There, most rivers are forced to flow eastwards parallel to fold axis not only due to lateral lengthening of folds during their growth (Ramsey et al., 2008) but also in relation with deformation along a set of oblique transpressive ridges: the Surmeh–Karebass and the Sabz–Pushan faults segments. In the central Fars, flat intermontane depressions form pounded areas filled by Quaternary sediments (e.g., Walker et al., 2011). These features appear characteristics of the plateau expansion into the Zagros belt (Khadivi et al., 2012). The lengthening and merging of forced folds promote river captures and provide a mechanism for increasing the river stream power at the origin of the observed transverse course of the Mand river (Ramsey et al., 2008). Oveisi et al. (2009) interestingly noticed, based on the analysis of river profiles that the most recent active uplift was not related to major active basement faults such as the MFF or Surmeh Fault but at distance from them. This suggests that erosion is currently mostly related to active thin-skinned deformation in this area.

A long term evolution of the drainage system in the northern part of the central Zagros has been proposed based on combined petrographic and detrital apatite FT thermochronometry study (Khadivi et al., 2012). This study showed that the Miocene sedimentary rocks were mainly sourced by the obducted ophiolitic complexes. This implies that this portion of the foreland was connected upstream, by transverse streams, to the High Zagros before ca. 15 Ma. The current river network, however, is no more drained by the HZ region where ophiolitic units are cropping out. This could be explained by the adjustment of river courses to the development of folding in the past 15 Myr. The south-westwards propagation of deformation caused by the widening of the Zagros–Iranian Plateau region likely after 12.4 Ma was probably too rapid for rivers to keep pace with fold uplift, hence preventing incision by antecedence or superposition. In the Lorestan,

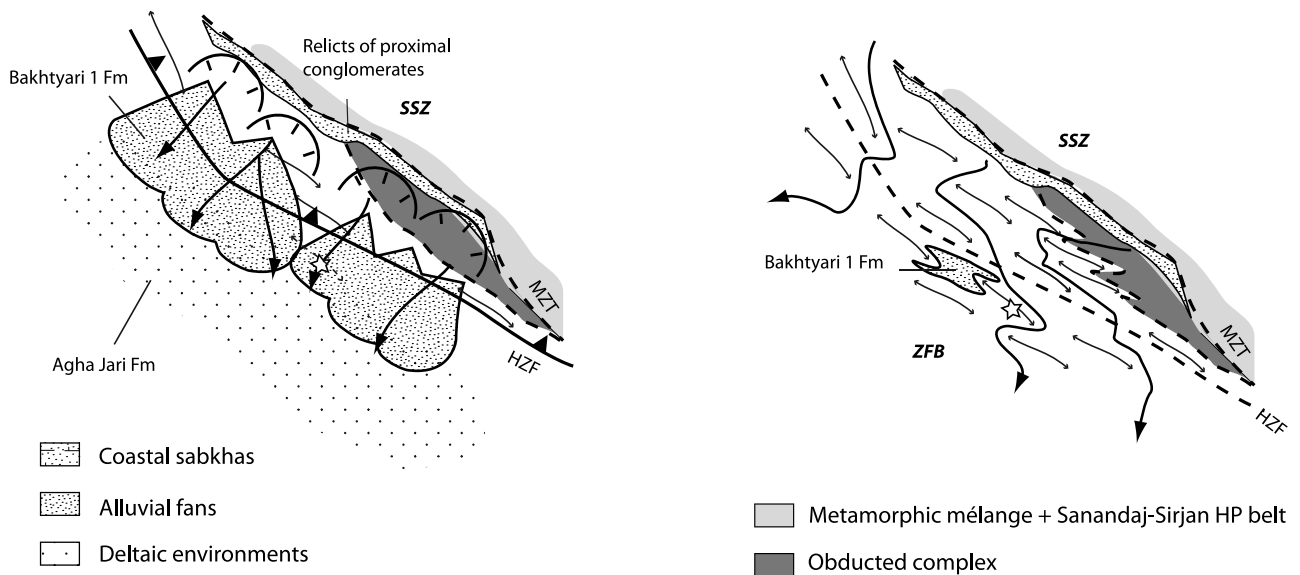
### A) Current fold/stream relationships



### B) Drainage reorganisation from transverse to axial river network

Early-Middle Miocene (16.6–13.8 Ma)

Late Miocene (<12.4 Ma)



**Fig. 6.** Main drainage basins and river networks of the Zagros folded belt. A) The Rud-e Karun catchment of the Lorestan region appears dominated by transverse rivers (152,000 km<sup>2</sup>) and is characterized by river gorges cutting through anticlines. By contrast, the Rud-e Mand catchment of the Fars region (80,000 km<sup>2</sup>) is characterized by axial rivers. B) Temporal constraints on the reorganization of the river network in the Fars arc from transverse to axial, in association with the elongation of folds and regional uplift after [Khadivi et al. \(2012\)](#).

the lower Agha Jari fluvial system flowed longitudinally in the foreland depression towards the ancestral Persian Gulf before being uplifted above the Mountain Frontal Flexure in latest Miocene times ([Vergés,](#)

[2007](#)). The uplift of this large region produced the shift of the Agha Jari fluvial system to the SW (close to the present position of Tigris and Euphrates rivers) and the fluvial incision during the Pliocene.

## 6. Crustal rheology, mechanics of folding and the building of Zagros topography

### 6.1. Mechanics of Zagros folding

#### 6.1.1. Buckling versus fault-related folding

Understanding better the fold mechanics and kinematics is of particular interest in the Zagros where several giant oil and gas fields are trapped within these structures in carbonate reservoir rocks. Zagros folding was originally interpreted as detachment folds (Colman-Sadd, 1978). Recent field studies and seismic profiling have shed new lights on the structure of folds and in particular to which levels these folds are associated to thrust ramps. In the Dezful domain, shallow thrust-related folds are likely caused by an intermediate level of decoupling in the evaporitic horizon of the Miocene Gashsaran Formation, the so-called “Upper Mobile Group” (Casciello et al., 2009; O'Brien, 1957; Sherkati and Letouzey, 2004; Sherkati et al., 2006; Vergés et al., 2011a) or in the Mishan Formation in the southernmost Fars (Molinaro et al., 2004, 2005a,b). The presence of intermediate incompetent layers within the sedimentary succession and their control on folding style have been discussed by different authors (Blanc et al., 2003; Sepehr and Cosgrove, 2004; Sherkati and Letouzey, 2004). Short wavelength anticlines in the Oligo-Miocene Asmari Formation in the Izeh zone indicate that Pabdeh and Gurpi formations are efficient intermediate décollement levels in this area. In the southeast of the Izeh zone and parts of the northeast Dezful Embayment this role is played by the Albian shales of the Kazhdumi Formation (Sherkati et al., 2005). In this region, seismic profiles give evidence for reverse faults, cutting through the folded Jurassic Surmeh Formation (Carruba et al., 2006; Sherkati and Letouzey, 2004) implying a basal detachment in deeper structural levels, most likely in the Hormuz Formation. These faults, where imaged, are always steep and characterized by limited displacements. In the Lorestan region, the multiple detachment folding exhibits a broad variety of fold wavelengths ranging from about 4 to 16 km with few of them reaching ~40 km (Casciello et al., 2009; Farzipour-Saein et al., 2009; Vergés et al., 2011a). The Pusht-e Kuh arc (Figs. 1 and 2) also shows areas in which the Asmari folds display small wavelengths. These are attributed to the lateral increase in the thickness of the Amiran succession up to 1000 m, interpreted as a main incompetent unit (Casciello et al., 2009).

The largest and most typical symmetric or slightly asymmetric detachment folds are observed in the Fars arc. There, one major issue is whether cover folding resulted primarily from the growth of buckle folds, e.g., symmetric/asymmetric detachment folds (e.g., Schmalholz et al., 2002; Mitra, 2003) or from thrust-related folding. Few available seismic reflection profiles (Jahani et al., 2009) helped by outcrop-scale observations suggest that folding is not linked to major thrust ramps in this region cutting through the cover, except where active basement faults have been inferred (Fig. 2). In this region, folding is characterized by regular, quasi-periodic folds with a peak wavelength ranging between  $14 \pm 3$  km and  $16 \pm 5$  km (Molinaro et al., 2005a,b; Mouthereau et al., 2007b; Yamato et al., 2011) with axial lengths sometimes larger than 100 km.

This suggests that buckling of the sedimentary rocks is the primary mode of fold development in the central Fars. This type of folding requires significant thickness and competency contrasts to grow. However, these parameters have rarely been quantitatively investigated. As noticed by Mouthereau et al. (2007b) the observed fold wavelengths are not compatible with simple elastic or viscous models of growing buckle fold instabilities. A key observation that may resolve this issue lies on the existence of mechanical stratigraphy, which defines a vertical rheological layering of the sediment cover. Indeed, the presence of intermediate weak horizons e.g., the Triassic–Permian evaporitic layers (Dashtak, Kangan, Dalan or Faraghan formations), the Upper Jurassic evaporitic Hith Formation or the Upper Cretaceous Kazhdumi shales together can act to reduce the

effective viscosity of the sediment cover (Fig. 7). A new set of visco-elasto-plastic numerical models, accounting for the fine-scale rheological stratification have therefore been carried out (Yamato et al., 2011). The model assumes a viscous power-law overburden resting on top of a linear viscous salt layer. Numerical simulations show that the addition of weak layers is sufficient to build detachment folds with the observed wavelengths. The best fitting rheological parameters are a viscosity of  $10^{18}$  Pa s for evaporitic intervals, and an average low friction angle of  $5^\circ$ . The model further indicates a total shortening of 11% to 17.5% consistent with shortening reported in the Zagros (Alavi, 2007; Blanc et al., 2003; Carruba et al., 2006; McQuarrie, 2004; Molinaro et al., 2005a,b; Mouthereau et al., 2007a, b; Sherkati et al., 2006; Vergés et al., 2011b). Without such intermediate weak layers an unrealistically thick salt layer of about 8 km would be needed to develop the folds. The lack of large-scale thrust ramps associated with folding is particularly striking and is directly related to the addition of these intermediate weak layers. However, as predicted by numerical modeling, faulting may occur between intermediate weak layers consistently with the occurrence of cover earthquakes between intermediate décollement levels (Nissen et al., 2011). This model therefore provides a mechanical understanding of the lateral changes from fault-dominated to fold-dominated style of deformation in the Zagros and in fold-thrust belts more generally. In this aspect, the Zagros Mountains differ from other salt-based fold-and-thrust belts such as the Jura Mountains, where large-offset faults are continuous across the stratigraphic sequence (Simpson, 2009).

The model confirms that buckling is a mechanically viable process to build the symmetric folds in the Zagros. In the Fars region, it supports the observation of limited deformation gradient across the belt, as confirmed by the regionally constant peak differential stresses value recorded by folded limestone formations ( $40 \pm 15$  MPa; Lacombe et al., 2007). Furthermore, this model places new bounds for the timing at which folding may form indicating that all folds may have developed coevally over the entire ZFB in 5.5 Myr, though this result needs further confirmation. However, we notice that such timing does not account for the interactions between cover and basement deformation, which may have modified from time to time the rate of deformation. We anticipate that folding could have started earlier. Additional stratigraphic constraints, especially at the southern deformation front are, however, needed to confirm this hypothesis.

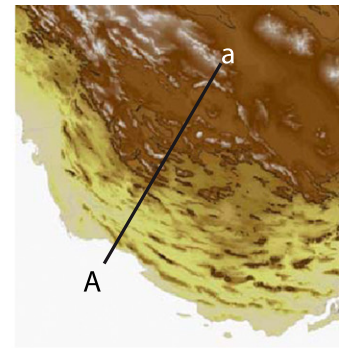
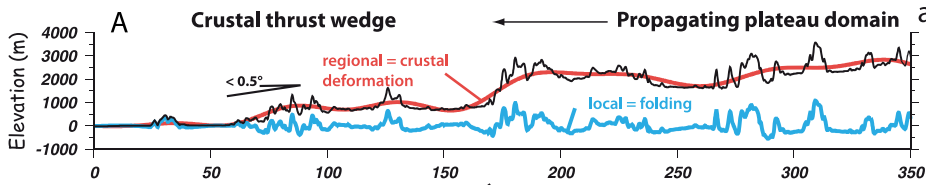
#### 6.1.2. Brittle deformation and Zagros folding

Fold-related fractures have originally been interpreted according to the structural domains of the fold in which they were observed (e.g., Bergbauer and Pollard, 2004; Erslev and Mayborn, 1997; Fischer et al., 1992; McQuillan, 1974; Srivastava and Engelder, 1990). In the view of using chronology of fracture sets, fracture patterns were related to fold kinematics (Anastasio et al., 1997; Bellahsen et al., 2006; Casini et al., 2011; Couzens and Dunne, 1994; Sanderson, 1982; Storti and Salvini, 2001; Tavani et al., 2006) and have been considered of key importance in the Zagros.

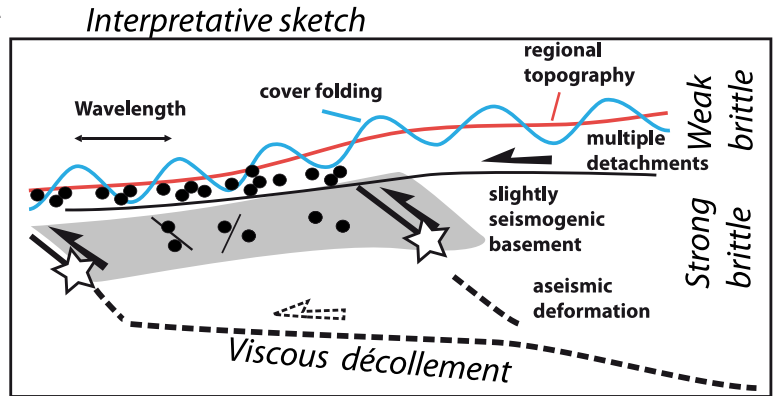
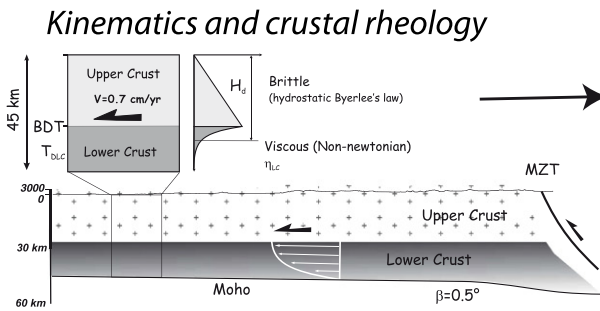
The Asmari Formation is an Oligocene–Early Miocene platform carbonate which is the most prolific oil reservoir in Iran and it is commonly regarded as a classic fractured carbonate reservoir, with production properties that depend strongly on the existence of fracture networks (e.g., Stephenson et al., 2007). In the Zagros, fracture studies were mainly reported from the Dezful, Izeh Zone and Lorestan areas. Some studies focused on the description of fracture sets within single anticlines, while others investigated fracture sets on a regional point of view. McQuillan (1973, 1974) demonstrated that many (if not all) regional fracture sets bear no relation to the geometry of folds formed during the Mio-Pliocene Zagros orogeny. This was later challenged by Gholipour (1998) who described that fractures in the Asmari reservoirs are associated with vertical and axial growth of concentric folding and therefore that they are mostly fold-related. In other places from the Dezful, both networks of diffuse fractures and



A) Wavelength components of the topography  
(Fars region, central Zagros)

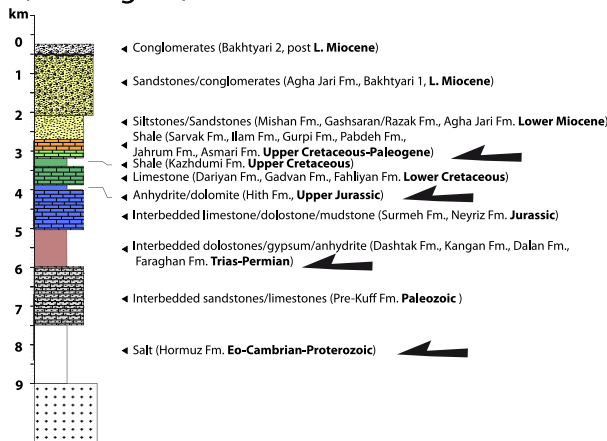


B) Crustal-scale orogenic wedge



C) Mechanism of cover folding

Stratigraphy and mechanical layering  
(Fars region)



Mechanical modelling of cover folding

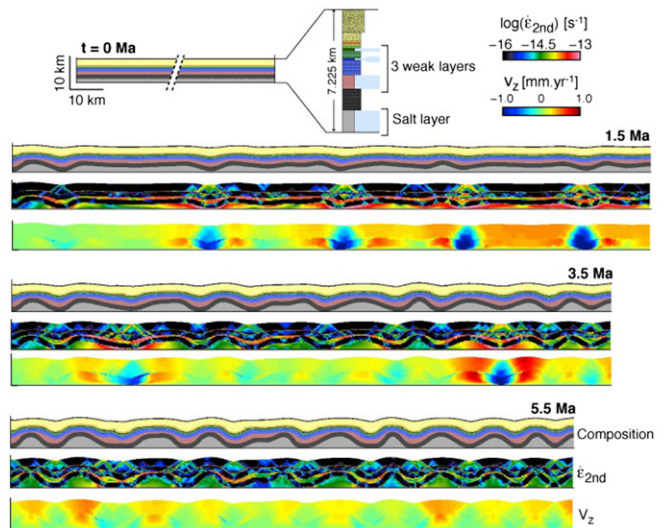


Fig. 7. Possible operating mechanisms of shortening in the cover and basement and the Zagros mountain building. A) Observed wavelength components of the topography showing the superimposition of regional topography (crustal deformation) and local fold topography (folding), modified after Mouthereau et al. (2006). B) Principles of the crustal-scale orogenic wedge modeling of the regional topography (topographic slope  $< 0.5^\circ$ ), modified after Mouthereau et al. (2006). C) Mechanism of shortening for the cover as inferred from numerical modeling (Yamato et al., 2011). Note the plastic/brittle shear zones in competent layers. The interpretative sketch shows the relationships between seismic deformation, crustal strength, main decoupling levels and the topography in the Zagros orogenic wedge.

fracture swarms (corridors) were recognized. In the Khaviz anticline, Wennberg et al. (2007) diffuse fractures fall into classical bed-perpendicular fold-related fractures sets. The density and height of these fractures in the Asmari Fm are controlled by the mechanical stratigraphy. In the Kuh-e Pahn anticline, fracture corridors striking parallel to the fold axis were interpreted as fold-related, but other well-developed fracture corridors recognized from satellite imagery,

clearly spatially unrelated to the detachment folding of the cover series, were interpreted as the distributed effect of deep-seated basement fault reactivation (Stephenson et al., 2007). Ahmadi et al. (2008) combined field structural observations and aerial/satellite image interpretation on several anticlines to propose a tectonic model highlighting the widespread development of pre-folding veins and other extensional micro/meso-structures in the Central

Zagros folded belt. Most joints/veins affecting the Asmari formation were found to predate the main folding episode. These early formed veins were reactivated (reopened and/or sheared) during Mio-Pliocene cover folding.

In the Fars, [Mobasher and Babaie \(2008\)](#) used remote sensing to analyze large-scale fracture systems and to relate them to either folding or strike-slip faulting along the Kazerun fault. A more recent study by [Lacombe et al. \(2011\)](#) has focused on a wide region across several anticlines from three structural domains in the Simply Folded Belt. In addition to classical fold-related fractures, these authors identified several pre/early-folding fracture patterns and used them to reconstruct the main compressional trends related to the early Zagros collisional history. The complex pattern of these pre-folding structures and the contrasting paleostress orientations through time in the different domains are related to the kinematics of both N–S and WNW-trending basement faults, above which cover was variably coupled during stress build-up and early deformation of the Arabian margin.

In the Lorestan, a regional analysis of the Cenomanian–Coniacian Sarvak and Ilam formations show pre-, syn- and post-folding fractures ([Casini et al., 2011](#)). Pre-folding fracturation consist of synsedimentary normal faults, systematic veins and stylolites. Syn-folding fractures correspond mainly to reactivation of previous fractures developing through-going fractures and reverse faults. Strike-slip faults typically cut through pre- and syn-folding structures and are probably related to fold tightening.

An important finding is that many fracture sets observed in the field are likely related to pre-folding tectonics, and that they were later reactivated during folding. Shearing and reopening of pre-existing vein/fracture sets appears finally to be a very important mechanism to control the small-scale brittle deformation within cover folds. Another important point is that the transmission of orogenic stress through the faulted crystalline basement of the Zagros was probably heterogeneous and complex, and that early basement block movements may have an impact on fracture development in the cover rocks. All these studies therefore draw attention on the need of carefully consider fractures unrelated to cover folding but likely linked to far-field orogenic stresses or to the distributed effect of deep-seated basement fault reactivation, before or after cover folding, to build more realistic conceptual fold-fracture models in the Zagros.

## 6.2. Evolution of stress patterns in the Zagros

[Authemayou et al. \(2006\)](#) focused on the Kazerun strike-slip segments, while [Lacombe et al. \(2006\)](#) reconstructed the stress field evolution from the Neogene to the present by means of inversion of fault slip data and earthquake focal mechanisms. [Navabpour et al. \(2007\)](#) reconstructed paleostress regimes in the High Zagros. These studies were completed by the analysis of calcite twin data ([Lacombe et al., 2007](#)) and of Anisotropy of Magnetic Susceptibility data ([Aubourg et al., 2010](#); [Bakhtari et al., 1998](#)).

In the western Fars, NE–SW and N020° compressional trends were identified during the Neogene and have been tentatively interpreted through different competing models. In the model invoking block rotations ([Bakhtari et al., 1998](#); [Lacombe et al., 2006](#)), the apparent change of the compressional trend from NE–SW to N020° through time is related to progressive vertical-axis clockwise rotations of cover blocks in relation to right-lateral wrench deformation accommodated by the Kazerun/Karebass/Sabz Pushan fault system. These rotations are consistent with the clockwise rotations suggested by AMS studies to explain the systematic anticlockwise obliquity between the pre-folding magnetic lineation and the fold axes (e.g., [Bakhtari et al., 1998](#)). Unfortunately, available paleomagnetic data show local inconsistent clockwise and anticlockwise rotations and are therefore currently unable to unambiguously support this model

([Aubourg et al., 2008](#)). In this scenario, the compression/shortening direction remained more or less N020° from the middle-late Miocene to present-day in this part of the Zagros. Alternatively, a regional counterclockwise rotation of the regional stress field during the Neogene has also been invoked ([Aubourg et al., 2010](#)). AMS data from Paleocene carbonates in the Simply Folded Belt record a N047° LPS during early to middle Miocene, while the middle to late Miocene clastic deposits recorded a N038° LPS prior to and during folding. The Plio-Quaternary paleostress trends are however consistently parallel to the present-day N020° shortening direction. In the absence of any block rotation or regional stress rotation, the early (pre-folding) local paleostress trends which deviate from the N020° trend has also been tentatively related to perturbations induced by underlying basement faults, depending on the domain considered and the degree of coupling between cover and basement above basement faults through the Hormuz salt layer ([Lacombe et al., 2011](#)).

In the Lorestan where folds are relatively rectilinear show small or non vertical axis rotation as demonstrated in the Changuleh anticline to the SW of the Mountain Frontal Flexure in which the data show 7° of clockwise rotation, which is within the angular error of the mean ([Homke et al., 2004](#)). In the Amiran anticline, center of the Lorestan, the mean direction yielded 351°, with an angular error of 5.9°, which represents a small but statistically significant counterclockwise rotation of 9° ([Homke et al., 2009](#)).

Miocene–Pliocene small-scale faulting observed in the cover indicate stress patterns (and regimes) nearly similar to those derived from the inversion of basement earthquakes, especially for the late N020° compression ([Lacombe et al., 2006](#)). Both types of data can thus be considered as reflecting the internal deformation of a pre-fractured crust as a whole (basement + cover), despite the occurrence of the thick Hormuz salt layer at the base of the cover. Regardless local complexities likely related to basement faulting, the N020° compressional trend agrees well with the current compressional trend revealed by inversion of the focal mechanisms of basement (and of few cover) earthquakes. It is also consistent with the geodetic shortening axis ([Walpersdorf et al., 2006](#)). This implies that the regional compression was approximately constant in space (across the Zagros collision zone) and time (during the late Neogene), in agreement with the stability of the Arabia–Eurasia convergence over the last 20 Ma ([McQuarrie et al., 2003](#)).

A compressional/strike-slip stress regime prevailed in the Zagros during the late Neogene, both in the cover and the basement. This regime accounts for the kinematics of the major faults ([Berberian, 1995](#)) and for the combination of strike-slip and thrust-type focal mechanisms of earthquakes whatever their magnitudes and focal depths. To a first-order, both the stress field and the deformation pattern therefore remained unchanged in the Zagros. Long-term AMS, calcite twin and fault slip data and short-term earthquake and GPS data are consistent with the idea that in the Fars, the Arabia–Eurasia convergence has been accommodated by both across-strike shortening and strike-slip faulting throughout the cover and the basement, with a minor component of belt-parallel extension.

It is worth noting that the Hormuz décollement poorly decouples principal stress orientations in the cover and the basement, although the GPS strain rate is much higher than the seismic strain rate ([Masson et al., 2004](#)). The present comparison of the stress field above and below the décollement of the still active Zagros belt yields a potential analog for ancient, salt-based fold belts.

A nearly similar conclusion was reached by [Navabpour et al. \(2008\)](#) in the Kermanschah area: stress tensor inversion of the earthquake focal mechanisms highlighted the partitioning in the W-Zagros with a N–S compression parallel to the plate convergence trend along the MRF and a NE–SW compression perpendicular to the fold axes across the ZFB. The integration of the recent compressive stress axes obtained from fault slip and seismic analyses on the shortening strain axes obtained from geodetic calculations revealed

different trends for the seismic compressive axis and the geodetic shortening axis, which suggests different degrees of partitioning within the basement and the sedimentary cover. Regardless the differences observed between deformation of the basement and sedimentary cover, the recent compressive stress directions seem to be similar throughout the basement and the sedimentary cover.

### 6.3. Mechanics of the Zagros orogenic wedge

The mechanics of accretionary wedges and fold-thrust belts can be described to first-order by the critical wedge theory (Chapple, 1978; Dahlen, 1990; Davis et al., 1983; Stockmal et al., 2007). A weak décollement will result in a thin, broad wedge, whereas a stronger décollement will form a steeper, narrower wedge. The ZFB has been viewed, for years, as a salt-based wedge (Davis and Engelder, 1985) in which the low topographic slopes of less than 1° (Ford, 2004; Mouthereau et al., 2006; Vergés et al., 2011b) reflect the low friction décollement. This interpretation has however been challenged by models showing that alone, such weak basal layer cannot support the topography (Mouthereau et al., 2006). It has been therefore proposed that the origin of the topography of the Zagros would be related to deeper, crustal-scale deformation (Fig. 7), and not solely to thrust imbrications in the sedimentary cover, in agreement with structural observations (Fig. 2). A way to achieve the building of a crustal-scale fold-thrust belt is to assume that the whole crust is sliding over the mostly aseismic lower crust (Nissen et al., 2011), strong enough to resist gravitational forces, that result from the thickening of the Arabian crust and the adjacent Iranian plateau. Indeed, the analysis of the topography shows that the Iranian plateau is currently expanding into the ZFB, as inferred by the plateau-like region in the northern ZFB and High Zagros of the Fars arc (Fig. 7a). This domain is bounded to the south by active deep-seated basement faults, thus forming the inferred critical wedge. A critically tapered wedge approach, applied to the upper brittle crust that is detached in a viscous lower crust, has been carried out and successfully applied. Based on this work, it can be proposed that the southern Zagros resulted from the propagation and stacking of deep-reverse faults rooting at depth into the middle-lower crust. We note that the applicability of the critical wedge approach requires that the internal part of the accretionary prism is at regional failure. This is indicated by the intense microseismicity in the basement (Tatar et al., 2004) and some large earthquakes (Mw ~6.7) that ruptured both the cover and the basement (Nissen et al., 2011). With respect to the cover, fewer basement earthquakes indicate that slip occurs on minor faults embedded in a stronger/elastic crust (Fig. 7). The lack of earthquakes in the lower crust, instead, likely reflects the decoupling in the aseismic ductile lower crust. This either reveals ductile shear in viscous lower crust as suggested by modeling or aseismic stable sliding in a brittle lower crust. Here, more studies and modeling are needed to decipher between the two hypotheses. Long-term ductile deformation at lower crustal levels is compatible with deformation of mylonite rocks at high pressure and temperature observed in hinterland of orogens or inferred from interseismic strain associated with major earthquakes (e.g., Bürgmann and Dresen, 2008, for the rheology of the lower crust). We estimate long-term (steady-state) effective viscosities as high as  $\sim 5 \times 10^{21}$  Pa s for the viscous lower crust able to sustain the Zagros topography and a 40 km thick deformed Arabian crust.

This model may also apply to the Lorestan area, although the timing of thrusting has probably been different. In the Dezful embayment, thick-skinned deformation is restricted within a narrower domain in the North and associated with a higher topographic slope, which could indicate a stronger basement. We speculate that the orientation of inherited structures in the irregular Arabian margin, coupled with the variable degree of convergence partitioning, could explain the along-strike variations in the wedge shape.

## 7. Discussion: strain distribution, tectonic model of Zagros mountain building and plate driving forces

### 7.1. Distribution of shortening across the Zagros and accommodation of Cenozoic Arabia/Eurasia convergence

In the following, we focus on how constraints summarized above may be integrated in the overall evolution of Arabia/Eurasia convergence. A significant effort is made to place quantitative estimates of finite shortening and deformation rates.

#### 7.1.1. Present-day slip rates and finite shortening in the Zagros since 22 Ma

GPS velocities show present-day convergence rates between Arabia and Eurasia of 19–23 mm/yr (McClusky et al., 2003). 7–10 mm/yr are absorbed in the Zagros Folded Belt (Hessami et al., 2006; Nilforoushan et al., 2003; Tatar et al., 2002; Vernant et al., 2004) most of which being accumulated at the Mountain Front Fault of the Fars arc (Oveisi et al., 2009; Walpersdorf et al., 2006).

Long-term shortening derived from balanced cross-sections across the Zagros Folded Belt and the High Zagros account for 50–70 km (Fig. 8; Blanc et al., 2003; McQuarrie, 2004; Molinaro et al., 2005a; Mouthereau et al., 2007b; Sherkat and Letouzey, 2004), which would be achieved in ~5–10 Myr to be consistent with the current shortening rates. A recent re-interpretation of the High Zagros thrust imbricates, in NW Zagros, accounting for the extreme thinning of the distal Arabian margin, led Vergés et al. (2011b) to determine a total shortening of 149 to 180 km across the ZFB since Late Cretaceous times (Fig. 2). Following field observations by Wrobel-Daveau et al. (2010), this interpretation places the true Neo-tethyan oceanic plate suture between the Bisotun block and the Gaveh-Rud forearc domain (Fig. 2). In the same issue, Mouthereau (2011) re-evaluated shortening from the High Zagros of the northern Fars region (Agard et al., 2005) by taking into account deformation related to the piling up of the Kermanshah ophiolitic units and obtained a shortening of 135 km since the early Miocene. Although quantitatively consistent, the latter interpretation places the collisional deformation within the plate suture, which was already tectonically overlying the Arabian margin above the MZT as the collision started. This occurred during suture tightening, when the Arabian crust was underthrust below the SSZ, and decoupled from the suture domain between ~35 Ma and ~20 Ma.

In the Zagros Folded Belt, cross-sections presented in Fig. 2, show consistent shortening of 15 km (Fars) and 21 km (Lorestan). The extrapolation of the current rates of arc-normal shortening of 8 mm/yr, across the ZFB, in the Fars (e.g., Walpersdorf et al., 2006) gives an initiation of shortening at ~2 Ma. In the Lorestan, where the arc-normal component is lower than ~2 mm/yr, the same approach constrains the initiation of shortening at ~10 Ma. Such a difference along the strike of the belt is unlikely. This reflects the fact that the current kinematic complexities, related to the seismic cycle, could not be resolved with available geological constraints, therefore highlighting the need for more stratigraphic constraints.

Based on comparison between short-term geodetic data and long-term geological constraints from regions surrounding the Arabia/Eurasia collision, Allen et al. (2004) inferred that the main episode of crustal thickening in the Zagros could have started 5–7 Myr ago. Temporal constraints, summarized in the study, established instead that contraction in the Zagros Folded Belt accelerated after 15–10 Ma. However, basement thrusting starting at 5.5 Ma, at the front of the Lorestan (Emami et al., 2010), likely reveals collisional tightening, during the Pliocene.

#### 7.1.2. Temporal distribution of collisional shortening in Arabia/Eurasia convergence: linkage between Zagros mountain building and Iranian plateau expansion

A synthesis of GPS data (ArRajehi et al., 2010), combined with reconstruction of past plate motion (McQuarrie et al., 2003), shows

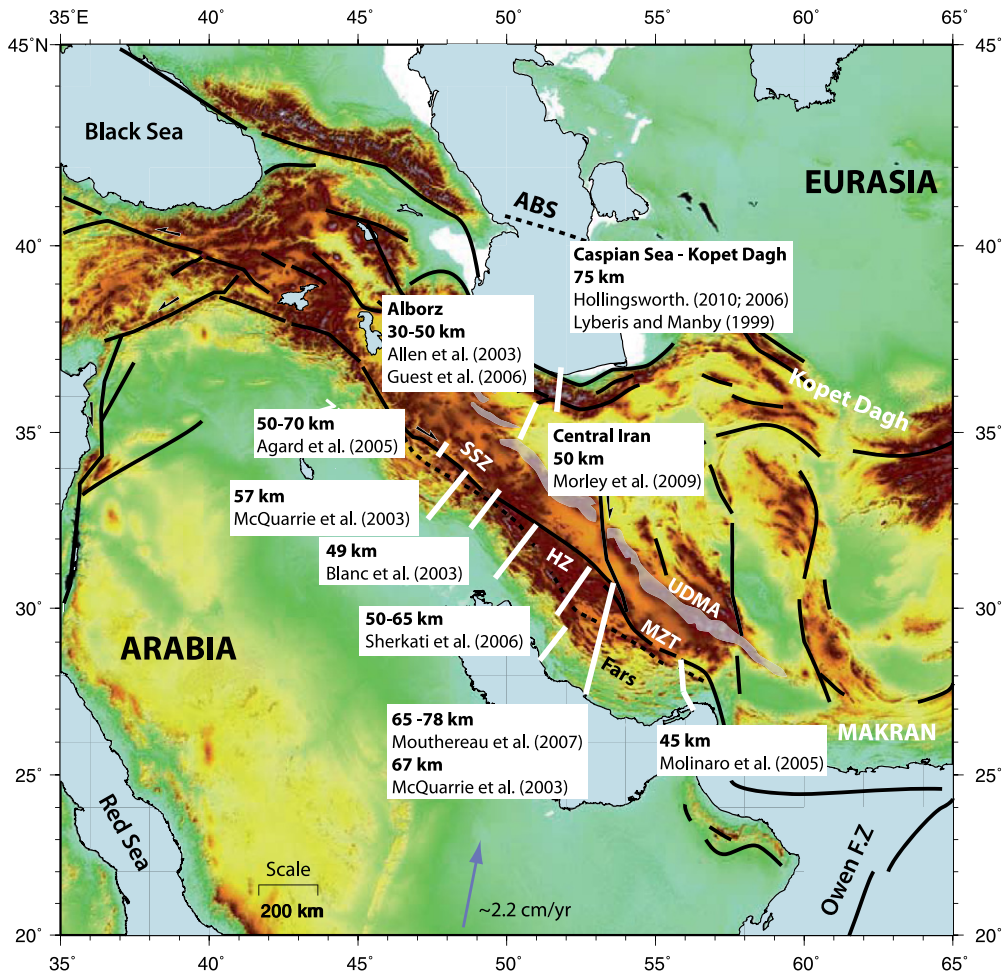


Fig. 8. Distribution of long-term shortening in the Arabia/Eurasia convergence estimated from balanced cross-sections.

that, within uncertainties, the AR/EUR convergence occurred at a rate of  $\sim 20$  km/Myr (Hatzfeld et al., 2003; Nilforoushan et al., 2003; Tatar et al., 2002; Vernant et al., 2004) since at least 22 Ma. A total convergence of 440 km must therefore be accommodated across the Zagros, SSZ, Central Iran, the Alborz and the Kopeh Dagh, by N–S distributed thickening, subduction/underthrusting and/or by strike-slip faulting.

By assuming conservation of mass and in-plane deformation Mouthereau (2011) used constraints on arc-normal shortening to derive the evolution of the crustal thickening and elevation since the early Miocene (Table 1). Among the 135 km to be accommodated in the central Zagros, about 15 km of shortening is accounted for by deformation in the ZFB after 15–10 Ma. Before ca. 15 Ma, shortening was accumulated in the High Zagros (50 km) through deformation of the distal margin (duplexing) and a component of underthrusting (70 km). A total shortening of  $\sim 120$ –180 km is reconstructed over the entire length of the Iranian plateau (300–450 km). There, crustal shortening is thought to have propagated into the SSZ as attested by deformation of the Neogene rocks in the south-central Iran and throughout the Iranian continental interior since  $\sim 20$  Ma, onwards. Deformation in the SSZ could therefore document a retro-arc thrust belt but more temporal constraints are, however, needed to confirm this. The onset of shortening in the Alborz Mountains occurred between 20 and 17.5 Ma, according to acceleration of accumulation rates (Ballato et al., 2008, 2011). In the Western Alborz, a rapid exhumational event is recorded at 12 Ma (Guest et al., 2006b) and was continuing until 6–4 Ma in Central Alborz (Axen et al., 2001) as indicating by low-temperature thermochronology. Shortening associated with the subduction of the south Caspian Sea at the Apsheron Sill is

constrained by the depth of earthquakes of 80 km (Jackson et al., 2002). Considering uncertainties on the timing of subduction initiation a value of  $\sim 75$  km satisfies both the data and the total convergence of 440 km (Table 1).

Overall, these results indicate that deformation originally localized near the plate suture (underthrusting and accretion of the Arabian crust) since  $\sim 25$  Ma, then propagated after 15–10 Ma both southward in the Zagros Folded Belt (Gavillot et al., 2010; Khadivi et al., 2010, 2012) and northward in the Alborz (Guest et al., 2006b), the Kopeh-Dagh, the south Caspian domain (Hollingsworth et al., 2010; Shabanian et al., 2009a,b) and in Central Iran (Morley et al., 2009) in association with the uplift of the Iranian plateau. Rapid exhumation in the Central Alborz at  $\sim 5$  Ma (Axen et al., 2001) and the coeval onset of increasing accumulation rates in the south Caspian Sea at 5.5 Ma (Allen et al., 2002) support this general trend (Fig. 8).

To account for current elevation, the mass balance model indicates that the crustal thickness of the Iranian interior (UDMA, Central Iran) was initially (prior to  $\sim 15$ –10 Ma) thinner than the surrounding Zagros and Alborz regions (Table 1). The originally thin crust characterizing Central Iran likely reflects the former backarc extension related to the Neotethyan slab rollback during the Eocene–Oligocene (Brunet et al., 2003; Ballato et al., 2011; Moritz et al., 2006; Morley et al., 2009; Verdel et al., 2007, 2011; Vincent et al., 2005). The Iranian lithosphere was consequently relatively weak as contraction initiated and hence shortened for low deviatoric stresses causing the inversion of extensional basins. As the crust of Central Iran progressively thickens, the forces necessary to balance the increase of potential energy associated with plateau growth also promoted deformation/

**Table 1**

Strain distribution in time and space in the Zagros and surrounding mountain belts of the Arabia/Eurasia collision. Constraints are from 1), Mouthereau (2011); 2) Khadivi et al. (2010, 2012); 3) Morley et al. (2009); 4) Allen et al. (2003) and Guest et al. (2006a,b); 5) Ballato et al. (2008, 2011), Guest et al. (2006a,b), Axen et al. (2001); 6) Homke et al. (2010), Khadivi et al. (2012) and this study; 7) Vergés et al. (2011b); 8) Brunet et al. (2003); 9) Jackson et al. (2002), Lyberis and Manby (1999); 10) Shabaniyan et al. (2009a,b); 11) Morley et al. (2009). + sign: extrapolated after shortening in central Iran. ++ sign: ZFB has been distinguished from the total shortening in the Zagros estimated from all other studies.

	Zagros Folded Belt	High Zagros	Sanandaj–Sirjan	Central Iran/UDMA	Alborz	South Caspian/Kopeh Dagh
Shortening	15–21 km++	128–159 km <sup>7</sup>	Unconstrained	38 km <sup>3</sup>	30–56 km <sup>4</sup>	75–80 km <sup>9</sup>
Model shortening <sup>1</sup>	15 km	50 km (duplexing) + 70 km (underthrusting)	130 km +	50 km	50 km	75 km
Model initial crustal thickness	40–45 km	45 to 5 km (thin distal margin)	35–40 km	32 km	> 32 km	30 km <sup>8</sup>
Onset of shortening	< 15 Ma <sup>2</sup>	25–22 Ma <sup>6</sup>	25–22 Ma <sup>6</sup>	10 Ma <sup>11</sup>	17–4 Ma <sup>5</sup>	< 10–3 Ma <sup>10</sup>
Current elevation	1–2 km	2–4 km	2 km	1.5 km	3–5 km	–0.5 km

uplift in the Alborz and the Zagros after 12 Ma. The current elevation in Iranian Plateau can therefore be explained by the original differences in the initial thickness of the continental crust inherited from subduction-related back-arc thinning events (Mouthereau, 2011). Additional effects due to 1) small-scale convective removal of Iranian lithospheric mantle during roll-back of an originally flat slab (e.g., Verdel et al., 2011) or 2) slab detachment (Jahangiri, 2007; Omrani et al., 2008) may have also contributed to the current regional elevation.

### 7.1.3. Partitioning of the Eurasia/Arabia convergence

While distributed thickening in the Arabia/Eurasia collision zone is revealed to have played a dominant role in building the regional topography, the ways in which the Cenozoic convergence has been accommodated through strike-slip faulting in the Zagros (e.g., Authemayou et al., 2009; Lacombe et al., 2006; Talebian and Jackson, 2004) and central Iran is debated (e.g., Allen et al., 2011; Meyer and Le Dortz, 2007; Walker and Jackson, 2004).

**7.1.3.1. Strike-slip faulting in the Zagros.** We have pointed out earlier that the kinematics of the Zagros collision is currently partitioning the N–S convergence, into a NW–SE orogen-parallel right-lateral strike-slip faulting along the MRF, and NE–SW orogen-normal shortening in the Zagros folds/thrusts (Authemayou et al., 2009; Talebian and Jackson, 2004). The kinematic role of right-lateral strike-slip faulting in the Zagros (Kazerun, Karebass, Sabz-Pushan and Sarvestan Fault) is diversely interpreted. There are viewed as faults bounding, counterclockwise rotated blocks, which accommodate an arc-parallel elongation between the partitioned domain of NW Zagros and no partitioned domain of the SE Zagros (Talebian and Jackson, 2004). This stretching, observed in the GPS data, is also emphasized by a component of belt-parallel extension, as recorded by fault slip data analysis, calcite twinning and active/quaternary faulting (e.g., Dasht-e-Arjan graben). Clockwise rotation between dextral strike-slip faults, as inferred from the analysis of stress reconstruction and evidence for primary fold curvature in the sedimentary cover (Bakhtari et al., 1998; Lacombe et al., 2006), however, do not support anticlockwise rotation of strike-slip faults. Alternatively, there are interpreted to reflect the distributed deformation at the termination of the MRF (Authemayou et al., 2006; Authemayou et al., 2009).

A cumulative dextral offset of ~50 km has been estimated along the MRF (Talebian and Jackson, 2002) based on the restoration of the drainage patterns and geological markers. It implies a shortening of ~50 km (or 70 km of N–S shortening), which is accounted for by the 149–180 km of shortening inferred from balanced cross-section in the NW Zagros (Vergés et al., 2011a,b). Taking into account the Quaternary slip rates calculated for the Main Recent Fault (MRF) and the Kazerun Fault (KF), Authemayou et al. (2009) concluded, like Talebian and Jackson (2002), that the motion along the MRF and KF has initiated 2–5 Myr ago. Assuming a roughly constant convergence rate of 20 mm/yr (ArRajehi et al., 2010; McQuarrie et al., 2003), it represents more than 70% of the total post-5 Ma N–S

convergence to be accommodated in the Arabia/Eurasia collision. This would therefore imply that the folding in the Zagros initiated at this time, which is refuted by the constraints on the timing of folding, uplift and exhumation (Fig. 2). Because these sets of active strike-slip faults in Iran could be connected to the North Anatolian Fault in Turkey, and are likely linked to the westward extrusion of Anatolia (Authemayou et al., 2009; Talebian and Jackson, 2002), the accurate dating of these faults is of major importance for the geodynamics of the region. We suggest that an age of ~10–12 Ma for the initiation of the MRF and KF would be in better agreement with the geological constraints in the Zagros and not in contradiction with the initiation of uplift of the Anatolian plateau at 10–11 Ma due to slab breakoff (Faccenna et al., 2006; Keskin, 2003; Sengor et al., 2003). The analyses of fault slip data (Ahmadhadi et al., 2007; Navabpour et al., 2008), consistently with plate reconstructions (McQuarrie et al., 2003), indicate that the convergence was oblique since 20 Ma. It should therefore be envisaged that the convergence was partitioned between a component of arc-normal contraction and a component of dextral strike-slip movement along a proto-MRF, which could have been positioned near the plate suture or within the upper plate (e.g., Sanandaj–Sirjan Zone).

**7.1.3.2. Strike-slip faulting in the Arabia/Eurasia collision zone.** There are several manners for deformation associated with strike-slip faulting to accommodate the plate convergence in continental interior (see review in Allen et al., 2011). They can rotate about vertical axes to accommodate N–S shortening and arc-parallel lengthening, as proposed for the Zagros (Talebian and Jackson, 2004). Alternatively, they can accommodate relative motion between non rotating blocks such as between central and eastern Iran (Deshir and Anar faults; Meyer and Le Dortz, 2007). These N–S right-lateral strike-slip faults absorb the differential displacement between the collision domain (central Iran) and the Makran subduction domain (eastern Iran) (Regard et al., 2010). A comparison between the long-term geological offsets and short-term geodetic displacement led Walker and Jackson (2004) to propose a cumulative N–S right-lateral shear in eastern Iran of 75–105 km. Assuming that the current kinematic configuration dates back to the inferred reorganization in the collision at 5–7 Ma (Allen et al., 2004), these authors postulated that the strike-slip faults of central Iran (e.g., Deshir and Anar faults) accommodate a small amount of shortening. However, Meyer and Le Dortz (2007) pointed out that a long-term kinematic model based on extrapolation of GPS rates may not be valid. They inferred that strike-slip faults of central Iran have accommodated a cumulative right-lateral shear of 90 km (Deshir and Anar faults) over the past 20 Myr. A paleoseismic study showed that the Deshir fault has been capable of producing earthquake as big as M~7 (Nazari et al., 2009). Therefore, there is evidence that right-lateral strike-slip faulting is not confined to the edges of the Lut block (Fig. 1). Taking into account the lack of current internal deformation across central Iran (Vernant et al., 2004) it has been suggested that deformation associated with strike-slip faulting

slowed in the last few Myr and shifted progressively to the east of the collision (Meyer et al., 2006; Allen et al., 2011).

The view that long-term N–S Arabia/Eurasia plate convergence in central Iran was partitioned by a combination of right-lateral strike-slip faulting and thrusting has been emphasized by Allen et al. (2011), analogous to the situation in the Zagros (Talebian and Jackson, 2004). In this aspect, the right-lateral strike-slip faulting is thought to accommodate the NW–SE orogen-parallel component of the N–S convergence. Thrusting that accommodates the NE–SW orogen-normal shortening was studied independently by Mouthereau (2011) to model the distribution of crustal thickening across the whole collision. These two models combined together provide a kinematic understanding of the way long-term N–S convergence is taken up in central Iran.

As noted earlier and consistently with Allen et al. (2004, 2011), deformation (strike-slip faulting and thrusting) seems to have slowed in central Iran. This was possibly related to a change in boundary conditions in the east, associated with the onset of the Afghan–India collision at ~5–2 Ma that stopped the possibility for lateral extrusion/escape. Here, in accord with the partitioning model of Allen et al. (2011), in which orogen-parallel lengthening is limited, we propose that the transition was progressive since 15–10 Ma and related to arc-normal thickening, which led to the uplift of the Iranian plateau and to the cessation of active deformation in central Iran. The progressive thickening in the Iranian plateau (e.g., SSZ) possibly promoted the decline of activity along NNW–SSE strike-slip faults like the Deshir or Anar faults. The progressive arc-normal shortening was contemporaneous with a late stage of westward lateral extrusion

that started after 10–11 Ma along the NAF in eastern Anatolia (e.g., Armijo et al., 1999) and in the Caspian Sea (Hollingsworth et al., 2006, 2010a).

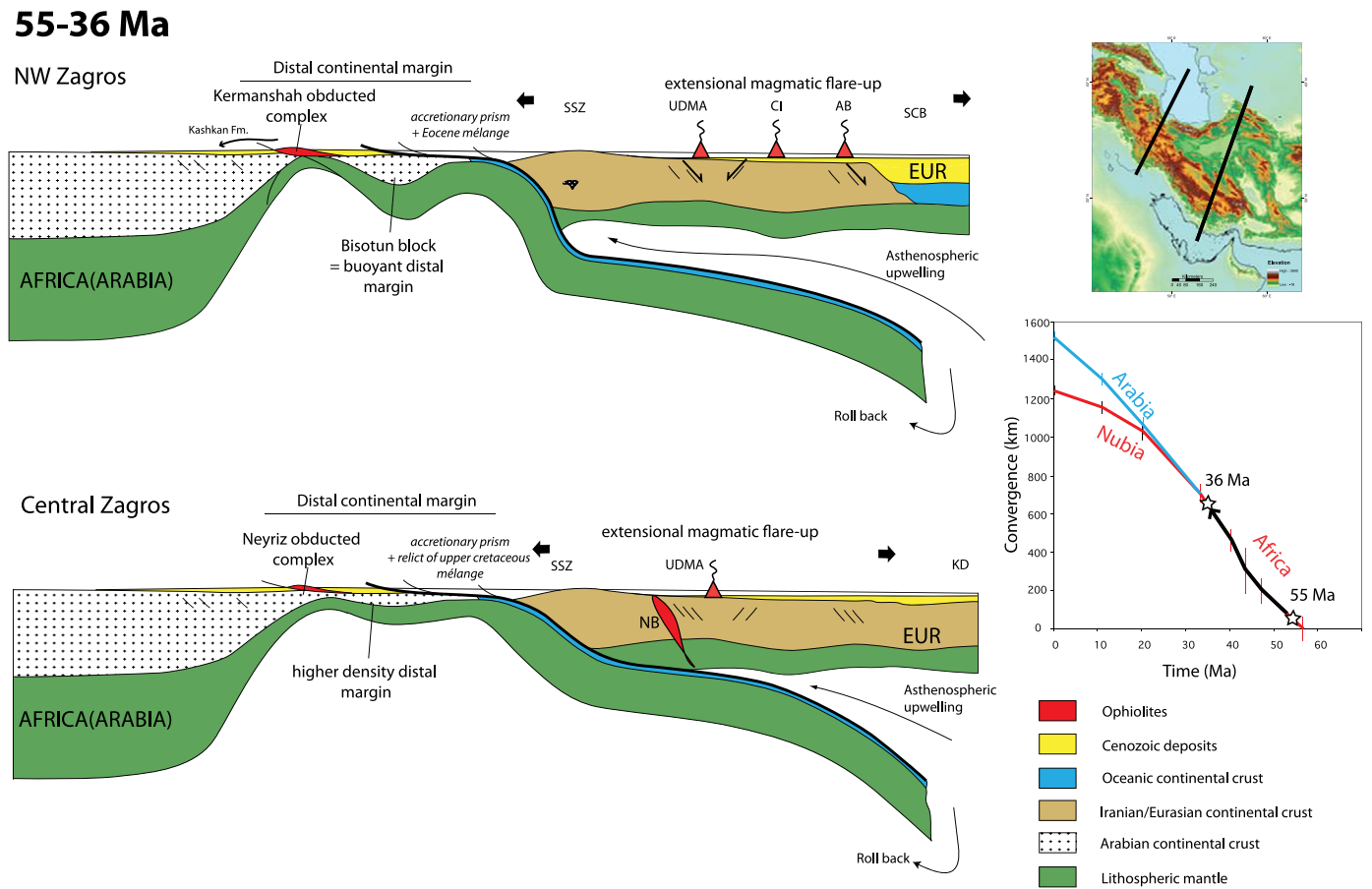
We conclude that the overall long-term distribution of deformation, a combination of partitioned strike-slip faulting and thrusting, in the Zagros collision region (we omit the yet subducting domain of eastern Iran) may be understood in the context of the N–S indentation of Arabia continent into Eurasia.

## 7.2. Stepwise tectonic model of the Zagros collision orogeny since 55 Ma

In the following, a stepwise tectonic model of the Zagros collision is proposed that emphasizes the role of the initial along-strike geometry of the Arabian continental margin. These differences are distinguished based on two possible scenarios that could possibly correspond to the NW Zagros and central Zagros.

### 7.2.1. Eocene stage (55–36 Ma): magmatic flare-up and initial Tethyan slab geometry

The subduction of the Neotethys beneath the Eurasian plate was characterized by Triassic–Jurassic Andean-like arc magmatism in the SSZ and has therefore been active at least 150 Myr prior to collision. In the Eocene, between 55 and 36 Ma a magmatic flare-up occurred in the UDMA and the Alborz region (Fig. 9; Agard et al., 2011; Berberian and Berberian, 1981; Brunet et al., 2003; McQuarrie et al., 2003; Verdel et al., 2011; Vincent et al., 2005). A constant plate velocity of 3 cm/yr is reported between 55 and 36 Ma (McQuarrie et al.,



**Fig. 9.** Lithospheric-scale cross-sections of NW Zagros and central Zagros at 55–36 Ma (subduction stage). Subduction of the Neo-Tethys and slab retreat trigger extension and magmatic flare-up in the Iranian plate. The differential along-strike stretching of the distal Arabian margin shows inherited buoyancy contrasts that had major impact on the subsequent accretion/subduction dynamics. Abbreviations: SSZ (Sanandaj–Sirjan Zone), UDMA (Urumieh–Dokhtar Magmatic Arc), CI (Central Iran), AB (Alborz), SCB (South Caspian Basin), KD (Kopeh Dagh), and NB (Nain–Baft basin).

2003), accounting for 570 km of convergence during this time interval.

Prior to flare-up, there is evidence that flat-slab subduction occurred in the mid-Cretaceous (Verdel et al., 2011) to account for widespread unconformity and deformation in the upper plate (Stocklin, 1968; Tillman et al., 1981) and the shift of magmatism northwards from the SSZ to the Alborz (Guest et al., 2006b). Flat subduction seems to have persisted until the Late Cretaceous/Paleocene as inferred by deformation, cooling (K-Feldspar Ar/Ar dating) and erosion in the Alborz (Guest et al., 2006a,b). The change in subduction angle, following the earlier suggestion by Berberian and Berberian (1981), likely resulted from a change in the slab density caused, for instance, by the subduction of buoyant slab segments, as argued from the Andean example (e.g., Martinod et al., 2010).

In the Zagros, the Late Cretaceous–Late Paleocene interval (65–55 Ma) is characterized by the migration of depocenters within the Amiran basin, which was fed by the erosion of ophiolitic complex (Kermanshah). This argues for a prolonged tectonic event and mountain building (Homke et al., 2009; Saura et al., 2011), a few Myr after the obduction episode.

It has been also argued that the initial flat-slab subduction has been followed by slab roll-back to explain exhumation of HP rocks in the Zagros and opening of small back-arc domain (e.g., Nain–Baft) at 90–65 Ma (Agard et al., 2006, 2011; Ghazi et al., 2011; Moghadam et al., 2010; Stampfli and Borel, 2002) in agreement with the slowdown of convergence at this time (e.g., Rosenbaum et al., 2002). Slab steepening (breakoff?; Agard et al., 2011) may have continued until the Paleocene–Eocene below SSZ to account for short-term and localized (Kermanshah) magmatism at 60–55 Ma.

In any case, tectonic reconstructions must account for the magmatic flare-up throughout central Iran that began at 55 Ma and lasted until 36 Ma. As noticed by Verdel et al. (2007), this is analogous to the initial flat-slab subduction during the mid-Tertiary magmatic flare-up in western US (Humphreys, 1995). Arc-like magmatism and extension characterized this period, which has been interpreted to reflect the onset of back-arc regime due to slab retreat (Ballato et al., 2011; Moritz et al., 2006; Morley et al., 2009; Verdel et al., 2007; Vincent et al., 2005). Alternatively, it has been suggested that back-arc extension and magmatism was accomplished by the “ablative” subduction without slab retreat, producing small-scale convective windows and partial delamination in the upper plate and melting (Agard et al., 2011) as proposed by Pope and Willett (1998) for the Andean plateau.

This period is associated with a remarkable stratigraphic hiatus in the Zagros that could possibly results from a moderate uplift of the Arabian continental margin, following the deposition of Kashkan red beds ca. 45 Ma and before the marine transgression at ~35 Ma (Fig. 3). At this time, the plate boundary was still located several 100 km to the north of the Arabian margin. A combined change in plate coupling due to increasing trench retreat and faster plate convergence with respect to late Cretaceous/Paleocene times, as inferred from Mesozoic plate reconstructions (Rosenbaum et al., 2002) could have led to compression in the lower plate, although this point needs more investigation.

As illustrated in Fig. 9, the distal Arabian margin at the end of the 55–36 Ma period, which is yet to collide with Eurasia, highlights along-strike differences in its initial geometry. While a buoyant continental block (Bisotun block) is present in the NW Zagros, it is absent in the central Zagros. Its buoyancy likely enhances slab retreat and magmatism in the SSZ of NW Zagros, while flat subduction enhances shortening of the Nain–Baft basin in central Zagros (Fig. 13).

#### 7.2.2. Oligocene stage (36–25 Ma): underthrusting of the Arabian margin

Back-arc extension continued in the early Oligocene (Fig. 10), but typical arc magmatism was terminated by ~36 Ma as indicated by geochemical analysis of the volcanoclastic Karaj Formation in the

Alborz (Ballato et al., 2011; Verdel et al., 2011). The petrological and geochemical composition of Early Oligocene basalts (33 Ma) points to an asthenospheric source, indicating the replacement (partial delamination) of the Iranian lithosphere by the asthenosphere. Convergence rates reduced from 3.1 to 2.4 cm/yr (McQuarrie et al., 2003). Assuming a convergence rate of 3 cm/yr, ~300 km has to be accommodated either by accretion or underthrusting between 35 Ma and 25 Ma.

The onset of underthrusting of the continental lithosphere of the distal Arabian margin beneath SSZ (e.g., Ballato et al., 2011) provides an explanation for both the convergence slowdown and increasing Tethyan slab retreat that promoted extension and asthenospheric upwelling below the upper plate. The driving mechanisms of underthrusting was probably the negative buoyancy of the distal Arabian continental margin. This is supported, for instance, by the extreme stretching of the distal Arabian margin, as inferred from the Kermanshah region (Wrobel-Daveau et al., 2010), which led to the removal of the buoyant continental crust and replacement by denser subcontinental mantle. As the continental lithosphere becomes denser than the underlying mantle underthrusting (subduction) of the continental lithosphere is promoted (e.g., Cloos, 1993). This is well-reproduced in analog experiments, showing that continental underthrusting is driven by slab-pull forces (Regard et al., 2003).

Underthrusting was probably dominant in early stage of the collision and plate convergence has to be accommodated at the plate suture. The main evidence for this relies on the timing of emplacement of fore-arc Gaveh-Rud domain (upper plate, southern SSZ) that yielded Rb–Sr age of ~34 Ma (Braud, 1987; Leterrier, 1985). This domain was thrust and folded together with older Lutetian–Bartonian flyschs (48–37 Ma) above the MZT, before deposition of the ~20–18 Ma Asmari Formation. Low-temperature apatite fission-track and (U–Th)/He thermochronological data, however, reveal rapid cooling only after ~25 Ma. The lack of evidence for exhumation in the Late Eocene likely indicates slow exhumation associated with the emplacement of thrust sheets in a marine Eocene accretionary prism. More generally, the onset of underthrusting of the Arabian continental margin implies a stronger plate coupling and the possibility of transferring stress in the upper plate. In Central Iran, this is indicated by the inversion of back-arc extensional basins (Fig. 13; Ballato et al., 2011; Morley et al., 2009; Verdel et al., 2007).

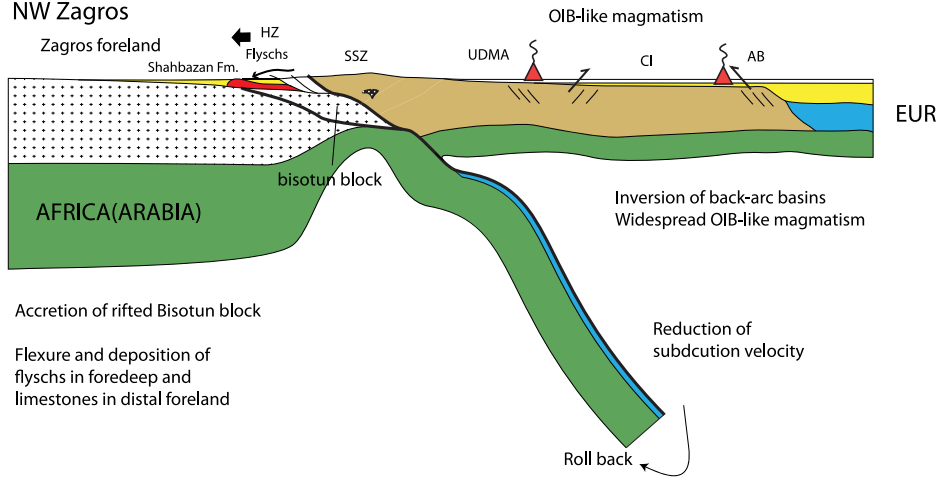
The deposition of Shahbazan carbonates at ~34 Ma (Homke et al., 2009) outline the onset of subsidence in the NW Zagros basin in relation to the premise of plate flexure. This may illustrate the onset of accretion of the buoyant Bisotun block onto the Arabian margin coevally with the emplacement of the forearc units. The accretion promoted the reduction in subduction velocity, thus increasing the Tethyan slab retreat in NW Zagros (Fig. 10). By contrast, the absence of the Bisotun rifted block in the central Zagros indicates more negative buoyancy that resulted in ongoing flat subduction.

#### 7.2.3. Early Miocene stage (25–15 Ma): thickening of Arabian margin and initiation of mountain building in the Zagros orogen

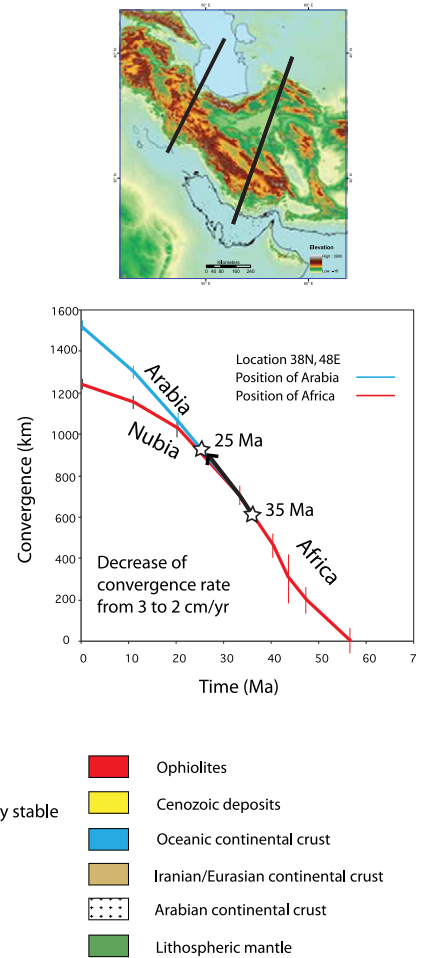
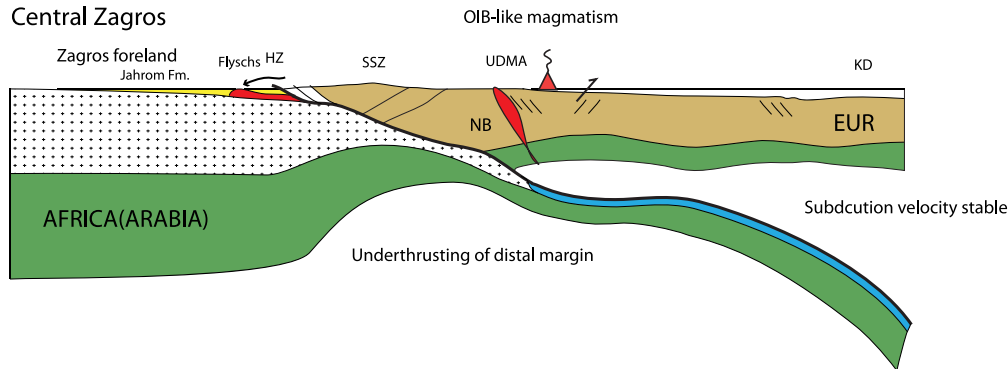
The early Miocene marks a significant change in the plate boundary conditions coeval with the onset of continental rifting of the Red Sea at ~24 Ma (Arrajehi et al., 2010; Chu and Gordon, 1998; McQuarrie et al., 2003) and propagation of Gulf of Aden opening westward since 18 Ma (Leroy et al., 2004). This induces a change in the direction of Arabia plate motion from NE to N at 25 Ma (McQuarrie et al., 2003). Among the 440 km of convergence to be accommodated since 22 Ma (assuming constant convergence of 2.2 cm/yr), about 70 km of collisional convergence were absorbed in less than 3 Myr in the central Zagros by ongoing underthrusting. In NW Zagros, roughly the same amount was taken up by the ongoing accretion of the Bisotun unit (Vergés et al., 2011a,b). After ca. 20 Ma (Figs. 11 and 13), both NW and central Zagros were shortened by accretion and thickening of the buoyant, normal thickness Arabian margin. This resulted in the reduction of subduction

## 36–25 Ma

### NW Zagros



### Central Zagros



**Fig. 10.** Lithospheric-scale cross-sections of NW Zagros and central Zagros at 36–25 Ma (onset of continental underthrusting). The negative buoyancy of the Arabian margin in central Zagros promotes underthrusting whereas accretion accelerates slab retreat in the NW Zagros. Accretion of Bisotun block is responsible for increasing foreland basin in the northern Zagros. Abbreviations: SSZ (Sanandaj–Sirjan Zone), UDMA (Urumieh–Dokhtar Magmatic Arc), CI (Central Iran), AB (Alborz), SCB (South Caspian Basin), KD (Kopeh Dagh), HZ (High Zagros), and NB (Nain–Baft basin).

(underthrusting) velocity below SSZ of central Zagros, thus increasing the roll-back of the flat slab.

The cessation of subduction-related magmatism in the UDMA occurred in the Oligocene–Miocene (~27 Ma) in NW Iran and was possibly delayed towards the southeast (from 16 Ma in Esfahan to 7 Ma near Kerman), as suggested from recent datings (see ages by Chung et al., 2010 presented in Agard et al., 2011) and in agreement with delayed slab retreat southwards.

Deposition of shallow-marine Asmari carbonates at ~20 Ma, unconformably overlying the older Shahbazan or Jahrom Formation to the north (Homke et al., 2009) and onlapping southwards the pre-collisional margin sediments (James and Wynd, 1965; Mouthereau et al., 2007b), outlines the initiation and migration of the plate deflection in the foreland. Uplift and exhumation occurred above the MZT, as confirmed by in-situ and detrital apatite fission-track thermochronologic analyses in the Lorestan and Fars regions (Homke et al., 2010; Khadivi et al., 2012). Magnetostratigraphic dating of sedimentary rocks hosting AFT grains cooled at 25 Ma, more accurately define that SSZ units came into contact with the Arabian margin at about 19.7 Ma (Khadivi et al., 2012).

Crustal shortening propagated northward from the SSZ throughout Central Iran after ~20 Ma (e.g., Morley et al., 2009). As pointed out earlier, this implies that a N-vergent retro-arc thrust belt developed in the SSZ in response to an increase in plate coupling. Together with the imbricate thrust stacking of the distal Arabian margin, they

form a double-sided orogenic wedge, typical of intracontinental like e.g. the Pyrenees (e.g., Beaumont et al., 2000). Shortening, uplift and exhumation reached the Alborz mountains at 20–17.5 Ma according to acceleration of accumulation rates (Ballato et al., 2008, 2011) and cooling ages of 17–15 Ma obtained from zircon helium data (Ballato et al., personal communication). Transition from marine to continental deposition in central Iran is outlined by the deposition of up to 7 km of clastic-dominated upper Red Formation dated at ~17 Ma (Ballato et al., 2008). Marine sedimentation dominated until ca. 15 Ma in the northern Zagros (Khadivi et al., 2010).

Stress build-up in the Arabian crust was responsible for the reactivation of inherited normal faults in the Precambrian basement and inversion of intramarginal basins. On the Iranian plate, subsidence occurred during the deposition of marine Qom Formation, accompanied by minor extension (28–18 Ma, Morley et al., 2009). This is interpreted to reflect thermal subsidence induced by cooling of the lithosphere, following the asthenospheric ascent and slab retreat. Alternatively, part of this subsidence may have originated from the plate flexure associated with tectonic loading produced by opposite vergence Alborz mountain belt and SSZ retro-arc thrust belt.

#### 7.2.4. Middle-Late Miocene (15–5 Ma): uplift of the Zagros folding and broadening of the Zagros orogen/Iranian plateau region

The oldest growth strata in the Zagros Folded Belt indicates that Zagros folding started at 15 Ma, coevally with the deposition of



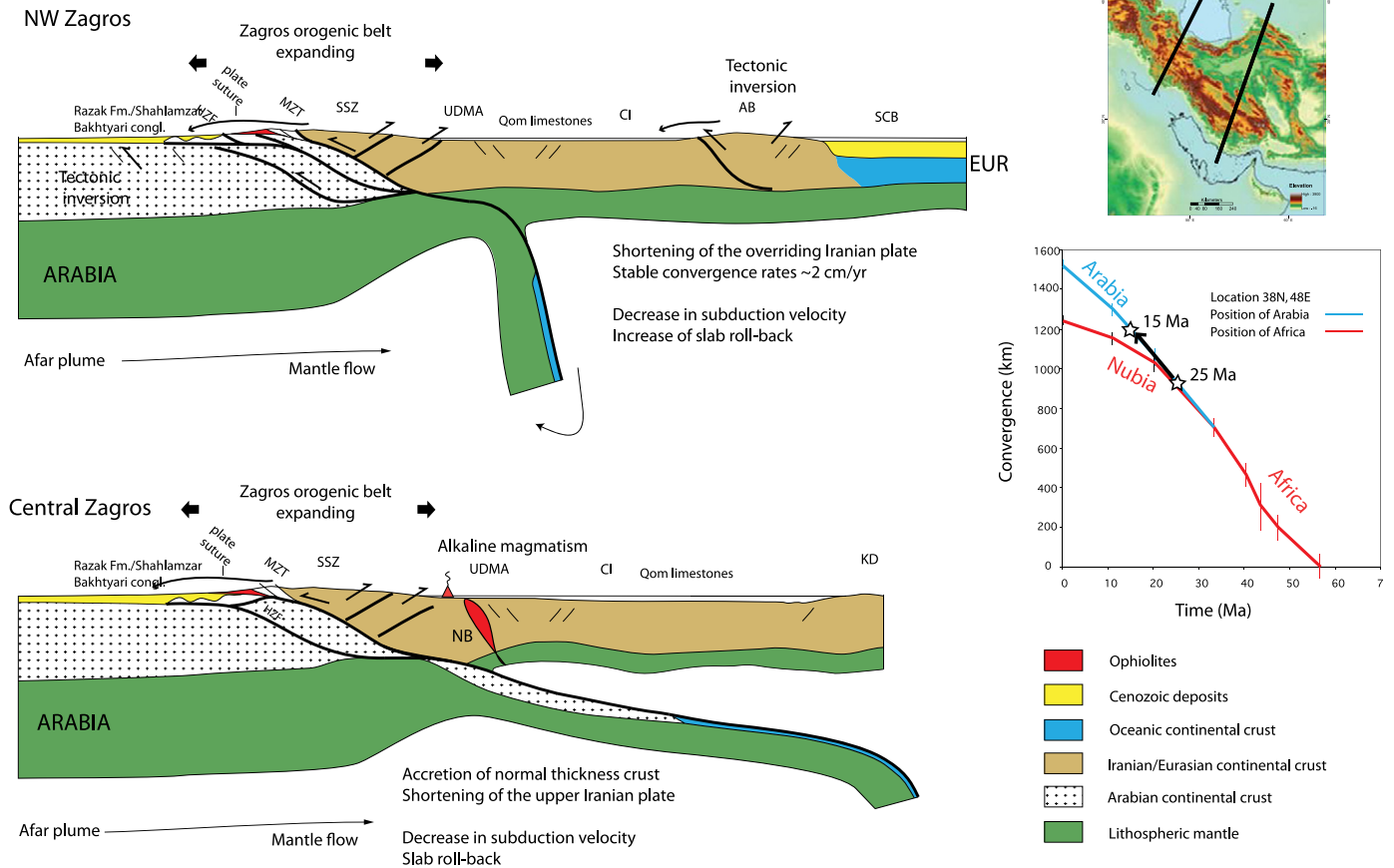
Bakhyari conglomerates (Khadivi et al., 2010) and propagated southward after 12 Ma (Emami, 2008; Khadivi et al., 2012; Mouthereau, 2011). This is in agreement with other magnetostratigraphic constraints in the northern Zagros (Izeh Zone) that support an 11 Ma age for folding (Emami, 2008). Moreover, apatite (U–Th)/He ages across the Lajin and the Dinar thrusts confirm that thrusting occurred between 19–15 Ma and 12–8 Ma in the High Zagros (Gavillot et al., 2010). This is also supported by thermochronometric dating with lower closure-temperature systems (apatite helium dating) that indicate ongoing (increasing?) exhumation at 12 Ma in the Zagros, tentatively correlated with deep-seated imbricate thrusting (underplating) (Wrobel-Daveau et al., 2011).

As suggested by the age constraints in the Zagros, basement thrusting and cover folding developed across the Zagros belt, and reached the current thrust front, leading to possible fold amplification at ~5 Ma (e.g., Mountain Front Fault) (Emami et al., 2010; Mouthereau et al., 2007b; Wrobel-Daveau et al., 2011). Crustal shortening and exhumation also migrated to the north, after 10 Ma in Central Iran (Morley et al., 2009), between 12 and 6–4 Ma in the Alborz (Axen et al., 2001; Guest et al., 2006b) as inferred from increasing accumulation rates in the south Caspian Sea at 5.5 Ma (Allen et al., 2002). The subduction of the south Caspian Sea at the Apsheron Sill was probably active since the latest Miocene based on comparison between geodetic constraints (Masson et al., 2007) and the depth of subduction slab (Jackson et al., 2002). Overall, these data show that shortening migrated toward low-elevated, undeformed foreland areas. Interestingly, these regions include the originally thin crust of

Central Iran in which first continental deposition is attested after 17 Ma.

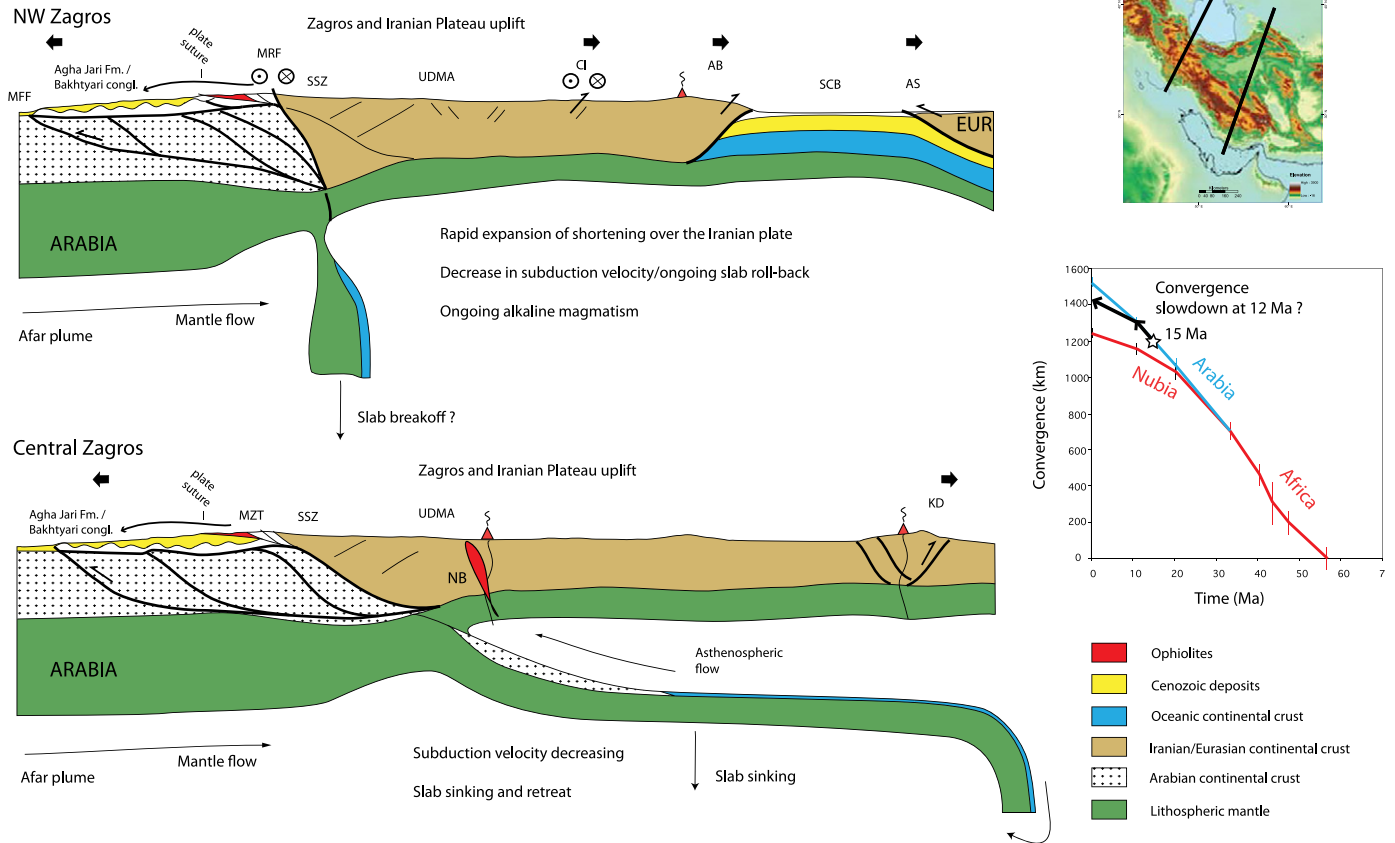
Taken together, these constraints show that uplift and exhumation began in the Arabia/Eurasia collision at 15–12 Ma (Figs. 11 and 12). The crustal shortening/thickening appears to be one of the main causes for the uplift of the Zagros/Iranian plateau region. Alternatively, breakoff of the Arabian slab after 10 Ma inferred from adakitic volcanism (Jahangiri, 2007) and gravity/thermal modeling in the central Zagros (Molinari et al., 2005a,b) has been proposed to be the main cause of plate coupling and shortening in the region (Omrani et al., 2008). However, neither kinematic reconstruction nor tomographic data confirm this possibility in the central Zagros. Indeed, as shown in Fig. 4, improved tomographic images display a shallow and flat subduction of the Arabian margin in this region over a distance up to 500 km below the Iranian plateau (Paul et al., 2010; Simmons et al., 2011). Underthrusting of the Arabian margin is, therefore, ongoing since at least the early Miocene. The lack of evidence for abrupt changes in slab geometry is consistent with a rather stable plate convergence (ArRajehi et al., 2010; Reilinger and McClusky, 2011). Furthermore, this is in line with the previous Eocene stage of flat-slab subduction inferred from the analysis of magmatism in the UDMA (Verdel et al., 2011). All together these observations rule out the hypothesis of a slab detachment after 10 Ma in the central Zagros (Figs. 12 and 13). We speculate that the southward migration of asthenospheric upwelling toward the plate suture, as proposed by Verdel et al. (2007), as the flat-slab region retreated, is perhaps responsible for the post-late Miocene adakitic

## 25–15 Ma



**Fig. 11.** Lithospheric-scale cross-sections of NW Zagros and central Zagros at 25–15 Ma (onset of crustal thickening). The continuous convergence gradually compressed and thickened the Arabian crust during this stage. Tectonic inversion of backarc basins starts and crustal thickening in the Zagros create plate flexure and the development of a foreland basin filled by clastic deposits eroded from the plate suture and SSZ. Abbreviations: SSZ (Sanandaj–Sirjan Zone), UDMA (Urumieh–Dokhtar Magmatic Arc), CI (Central Iran), AB (Alborz), SCB (South Caspian Basin), KD (Kopeh Dagh), HZ (High Zagros), NB (Nain–Baft basin), and MZT (Main Zagros Thrust).

## 15–5 Ma



**Fig. 12.** Lithospheric-scale cross-sections of NW Zagros and central Zagros at 15–5 Ma (uplift and expansion of the Zagros and Iranian plateau). The Iranian plateau has been thickened and tectonic progressively declines in this region. Topography and deformation expands towards low-elevated area of the Zagros, Alborz and Kopeh Dagh forelands. Subduction of the South Caspian basin starts at this time. Deformation in the Zagros through folding and basement shortening is more pronounced. The MRF initiates in response to the slab detachment in the eastern Anatolia, slab steepening in NW Zagros. Adakitic magmatism resumes in the central Zagros due to renewed Arabian slab retreat and e.g., mantle-derived melts at lower crustal levels. Abbreviations: SSZ (Sanandaj–Sirjan Zone), UDMA (Urumieh–Dokhtar Magmatic Arc), CI (Central Iran), AB (Alborz), SCB (South Caspian Basin), KD (Kopeh Dagh), HZ (High Zagros), NB (Nain–Baft basin), MZT (Main Zagros Thrust), and MRF (Main Recent Fault).

volcanism in the UDMA. Post-collision magmatism as young as ~2 Ma in the Kopeh Dagh (Shabanian et al., 2009a,b) may highlight a possible slab steepening/retreat at the nose of the underthrusting domain. This supports the hypothesis that adakites resulted from melting of a mafic Iranian lower crust (inherited from the previous magmatic flare-up stage) analogous to Miocene adakitic and K-rich magmas emplaced in the Tibetan crust (e.g., Guo et al., 2006; Guo et al., 2007). Tomography data support, however, slab detachment in NW Zagros in agreement with slab steepening and breakoff inferred in Anatolia (Keskin, 2003) and the propagation of the North Anatolian Fault (Authemayou et al., 2006; Faccenna et al., 2006; Regard et al., 2005) at ~10–11 Ma when the uplift of the Kars–Erzurum plateau initiated. Likewise Authemayou et al. (2009) and Talebian and Jackson (2002), we anticipate that the active right-lateral shear along the MRF may have been initiated in response to breakoff to the north and westward tectonic extrusion of Anatolia.

A post-Miocene tectonic reorganization at 3–5 Ma has been suggested by many authors (e.g., Allen et al., 2004; Authemayou et al., 2006; Axen et al., 2001; Shabanian et al., 2009b) and has been inferred from the drastic changes in the regional tectonic regime during Quaternary (e.g., Abbassi and Farbod, 2009; Javidfakhr et al., 2011; Regard et al., 2005; Ritz et al., 2006; Shabanian et al., 2010).

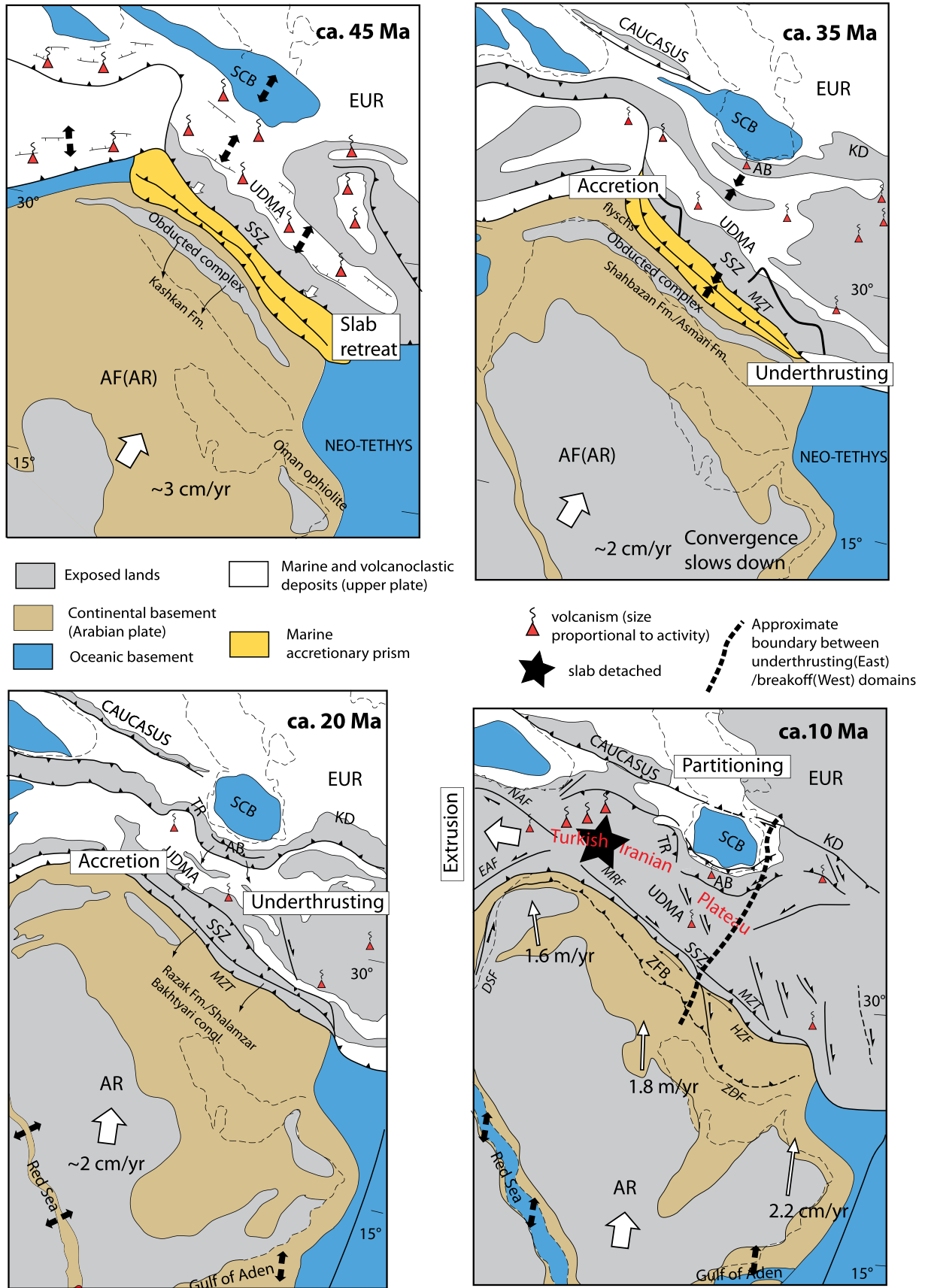
In the Zagros, a phase of fold tightening may have occurred at 5 Ma, as suggested by the unconformity of the most recent Bakhtyari Formation and amplification of basement thrust at the front, but there is no definitive observation to infer a rapid kinematic change at 5 Ma in the Zagros. In the Alborz, exhumation is seen to have amplified

recently as evidenced by two marked events at ~12 Ma and later at 6–4 Ma. In the Kopeh Dagh region, north of the Lut block (subduction related domain) deformation seems to have started after 10–5 Ma (Hollingsworth et al., 2006, 2010). This may provide evidence for the progressive transfer of collisional deformation northward into the continental interior, helped by underthrusting in central Zagros (or slab steepening/breakoff in NW Zagros) and associated plate coupling. As shown in this study, available time constraints indicate a progressive arc–normal thickening, uplift and westward extrusion beginning at 15–12 Ma (Fig. 11), which does requires a kinematic change at ~5 Ma for the entire Arabia/Eurasia collision.

We conclude that uplift and shortening in the Zagros–Iranian plateau may have resulted from increasing plate coupling through a combination of slab detachment in NW Zagros (Lorestan, west of 51°E) and continuous underthrusting in central Zagros (Fars, east of 52°E). Recent plate reconstructions have pointed out that the convergence may have reduced by 30% after 12 Ma (Austermann et al., in review). If correct, it is possible to envisage that the topographic load of the Zagros and Iranian plateau region has been the primary cause for the reduction of plate motion, in a way similar to prediction from the Andes (Iaffaldano et al., 2006).

### 7.3. Plate kinematics and the driving forces of plate convergence

After the initiation of continental underthrusting in the Zagros at ~35 Ma, the Arabia's plate velocity decreased by ~30%, from ~3.1 cm/yr to ~2.4 cm/yr (McQuarrie et al., 2003). The onset of



**Fig. 13.** Key tectonic events and plate kinematics illustrated on schematic paleotectonic reconstructions for the Middle East region, modified after Barrier and Vrielynck (2008). The increase of emerged lands illustrates the progressive uplift of the Zagros and Iranian plateau in relation with crustal thickening. A major change occurred after 20 Ma, most likely at 15–12 Ma, as shortening induced slab steepening breakoff at 10–11 Ma and a regional kinematic reorganization. Abbreviations are : SSZ (Sanandaj–Sirjan Zone), UDMA (Urumieh–Dokhtar Magmatic Arc), CI (Central Iran), AB (Alborz), SCB (South Caspian Basin), KD (Kopeh Dagh), HZ (High Zagros), NB (Nain–Baft basin), MZT (Main Zagros Thrust), MRF (Main Recent Fault), ZFB (Zagros Folded Belt), and ZDF (Zagros Deformation Front).

underthrusting of the continental lithosphere driven by the negative buoyancy of the distal Arabian margin provides an explanation for both the convergence slowdown and increasing Tethyan slab retreat (Fig. 9). After 25–20 Ma, the motion of the Arabian plate, as inferred from Euler poles and geological reconstructions remained roughly stable (ArRajehi et al., 2010; McQuarrie et al., 2003; Reilinger and McClusky, 2011). It has been therefore inferred that slab pull forces acting on the Arabian and Neo-tethyan oceanic slab were the controlling parameters of the kinematics of the Arabian plate (Bellahsen et al., 2003; McQuarrie et al., 2003).

The slab pull effect in the Zagros was probably enhanced by the negative buoyancy of the distal Arabian margin that resulted from the removal of the buoyant continental crust inherited from its Mesozoic stretching. This allows sustaining the stable Arabian plate motion against Eurasia and permitted ongoing underthrusting (subduction) of the continental lithosphere, analogous to the Himalaya (Capitanio et al., 2010). This mechanism has been mostly effective from 35 to 20 Ma and especially in central Zagros, as the convergence involved the thinnest part the Arabian margin (Fig. 10).

A number of successive tectonic events led to the detachment of the Arabian plate: at ~24 Ma, continental rifting initiated in the Red Sea (Arrajehi et al., 2010; Chu and Gordon, 1998; McQuarrie et al., 2003), during the Miocene at ~18 Ma when the Gulf of Aden opened (Leroy et al., 2004) and finally at ~11 Ma when full ocean spreading occurred (Reilinger and McClusky, 2011). Together with the development of the Afar plume and East African rifting at 20–15 Ma (Pik et al., 2008), these constraints suggest the increasing gravitational potential energy along the Red Sea and the Gulf of Aden. The ridge push effect, even limited could have contributed, with slab pull, to the northward indentation of Arabia against Eurasia (Becker and Faccenna, 2011; Bellahsen et al., 2003).

Constraints presented in this study support tectonic models in which the topography of the Zagros belt and uplift of the adjacent Iranian plateau were driven by progressive tightening of the collision, closure of back-arc basins in central Iran and widespread crustal thickening in the Iranian plateau after 15–12 Ma. An increase in shortening rates in the overriding plate has promoted a reduction of Arabian subduction velocity. About ~70% of the Arabian plate velocity was consumed by crustal thickening and strike-slip faulting in the overriding plate since ~20 Ma (Table 1 and Mouthereau, 2011). Thus, the long-term subduction velocity of the Arabian plate relative to overriding plate reduced from ~2 cm/yr after 35 Ma to ~1.3 cm/yr after 20 Ma. Slab steepening/breakoff as imaged by tomography in NW Zagros (Fig. 4; Chang et al., 2010; Simmons et al., 2011) has perhaps been promoted by such a decrease in subduction velocity. In the central Zagros, however, underthrusting of the Arabian plate seems to have continued during the whole collision. This may be explained by faster convergence velocity in the southeast and a shorter oceanic slab, due to collision obliquity, and anticlockwise movement of the Arabia plate relative to Eurasia.

Evidence for slab breakoff in NW Zagros effective active since ~11 Ma implies a reduction in the slab pull effect. Assuming a stable plate convergence over this period, it has been proposed that the main force driving plate motion is the mantle drag below the Arabian continent sustained by the sinking of the detached Tethyan slab in the lower mantle (e.g., Alvarez, 2010; Conrad and Lithgow-Bertelloni, 2004).

Recent improved plate reconstructions have pointed out that the convergence may have reduced by 30% after 12 Ma (Austermann et al., in review). This could have resulted from slab detachment in NW Zagros at this time or/and the increase of resistance at plate boundary caused by the increase in the regional topography. The present study indeed supports uplift, exhumation and shortening in the Zagros and Iranian plateau development starting at 15–12 Ma, coevally with slab detachment in eastern Turkey and the NW Zagros.

The opening of the Gulf of Aden at ~11 Ma contemporaneous with the increase in gravitational forces arising from the Zagros collision, which possibly reduced plate convergence, and slab breakoff at ~10–11 Ma, suggest that the role of the slab pull force in driving the Arabian plate motion was considerably reduced. We therefore speculate that heterogeneity in mantle density, as seen from mantle upwelling at the Afar plume and large-scale mantle flow is the main driver of the Arabia plate motion in agreement with recent findings based on models of global mantle circulations (Becker and Faccenna, 2011).

## 8. Conclusions

### 8.1. Timing of collision, onset of uplift/exhumation and mechanism of the Zagros mountain building

The Zagros collision sensu lato initiated at 35 Ma when the distal continental margin, driven by its negative buoyancy, was underthrust beneath the upper Iranian plate. The onset of crustal thickening and collision sensu stricto, started at ~25 Ma, as recorded by the coeval exhumational and foreland clastic deposition. However, uplift, deformation and exhumation across the Zagros, and throughout the Arabia/Eurasia collision zone, started later at 15–12 Ma. In the upper plate, the progressive shortening/thickening of the initially weak Iranian crust increased the topographic elevation and potential energy, leading to the cessation of contraction in the plateau region. Ongoing plate convergence promoted the expansion of the Iranian plateau towards low-elevated, undeformed areas, therefore focusing active shortening in forelands of the Arabia/Eurasian collision. Most of the current shortening is being accumulated at the Mountain Front Fault. Not excluding the contribution of slab breakoff in NW Zagros, evidence of distributed long-term crustal shortening over the all Arabia/Eurasia collision indicates that crustal thickening is one of the main causes for the uplift of the Zagros/Iranian plateau region.

### 8.2. Underthrusting/accretion and the role of margin inheritance in collisional shortening

The 55–35 Ma magmatic flare-up throughout central Iran resulted from the active flat-slab subduction. The onset of OIB-like magmatism in central Iran, at 35 Ma, reveals the upwelling of asthenospheric mantle, likely promoted by slab retreat and the reduction of plate convergence. From the recent tomography analyses of the Middle East, it is established that slab steepening or slab breakoff is ongoing in NW Zagros. In the south, however, improved resolutions of tomographic models favor a model of underthrusting of the Arabian plate beneath central Iran. The along-strike changes in subduction dynamics may emphasize the original difference in the buoyancy of the distal margin. In the NW Zagros, the onset of flexural subsidence suggests that the accretion of the rifted Bisotun continental block occurred between 35 and 25 Ma. This likely decreased the subduction velocity and promoted slab sinking. In the central Zagros, the lack of such a buoyant continental block enabled the continuous underthrusting of the Arabian plate. More structural and temporal constraints at the plate suture are however needed to confirm the proposed scenario.

### 8.3. Structure of the ZFB, topography and rheology of the Arabian crust

The structure of the Zagros fold belt is characterized by superimposed cover folding that is cut by deep-seated basement thrusts. Not to exclude other contributions, basement shortening is one of the main causes of the regional Zagros topography, as inferred from balanced cross-sections and mechanical models. The basement is, however, currently less seismogenic than the sedimentary cover, although few large earthquakes ( $M_w \sim 6.7$ ) can rupture both the cover

and the basement. Low geothermal gradient in the region is compatible with the presence of earthquakes in the lower crust, down to depths of 30 km, triggered by active faulting in the upper crust. Do these deformation patterns reflect a strong or a weak basement? Critical wedge approach applied to the Zagros favors the hypothesis of strong brittle crust detached above a weaker ductile lower crust. It is nevertheless likely that ductile thickening played a significant role, especially below the SSZ, where the crustal root reaches the depth of 70 km. In contrast, the sedimentary cover formed by a thick layered succession with multiple embedded detachment levels deformed by buckling.

#### 8.4. Collision kinematics and plate driving forces

After the initiation of continental underthrusting in the Zagros at ~35 Ma, the Arabia's plate velocity decreased by ~30% at 25–20 Ma and remained constant since then. Slab breakoff in NW Zagros was probably effective at ~11 Ma and implies a reduction in the slab pull effect. Moreover, recent plate reconstructions seem to indicate a reduction of plate convergence after 12 Ma. The opening of Gulf of Aden at ~11 Ma, contemporaneous with the increase in gravitational forces in the Zagros collision, and slab breakoff at ~10–11 Ma suggest that the role of the slab pull force in driving the Arabian plate motion was considerably reduced. We speculate that the mantle upwelling at the Afar plume and induced large-scale mantle flow is the main driver of the Arabia plate motion.

#### 8.5. Kinematic reorganization of the Arabia/Eurasia collision at 15–12 Ma?

Temporal constraints on deformation outline the regional mountain and plateau uplift after 15–12 Ma. At the same time ~10–11 Ma, break-off beneath eastern Anatolia led to the uplift of the Kars–Erzurum plateau. This occurs coevally with the initiation of westward tectonic extrusion along the North Anatolian Fault and resulted in right-lateral shear along the MRF in the Zagros. Before that date the obliquity of the convergence was accommodated by the partitioning of arc-normal shortening and arc-parallel strike-slip faulting. We speculate that the main kinematic change in the Zagros region occurred at 15–12 Ma, as the Zagros uplifted, therefore slightly before the Arabian slab detached beneath eastern Anatolia and NW Zagros. This change possibly induced a decrease in the plate convergence rate.

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