The long-term evolution of the Doruneh Fault region (Central Iran): A key to understanding the spatio-temporal tectonic evolution in the hinterland of the Zagros convergence zone

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A better understanding of intraplate deformation requires the knowledge of the space–time scales involved in its development and to decipher possible links with the dynamic evolution of the plate boundaries. Central Iran provides an ideal test site to approach this scientific issue, since it is characterised by a prolonged history of Mesozoic–Cenozoic intraplate deformation that has been interfering with the spatio-temporal re-organization of the Zagros convergence zone along the Eurasia plate boundary. This study focus on the Doruneh Fault (DF) region that is considered as the northern mechanical boundary of the Central East Iranian Microcontinent. By combining field investigations with apatite low-temperature thermochronology, we present a revised tectono-stratigraphic scenario for the DF region, typified by a punctuated history of fault-related exhumation, burial and cooling history back to the Upper Cretaceous. When framed at regional scale, these results attest that the Zagros convergence zone, and its hinterland domain were fully mechanically coupled since ca. 40–35 Ma, a time lapse that is here referred as to the onset of continental collision along the Arabia–Eurasia plate boundary. In this scenario, the DF region operated throughout the Cenozoic as a major zone of residual stress accommodation and transfer in the hinterland domain of the Zagros convergence zone. Results of this study also suggest that the tectonic evolution along the Arabia–Eurasia plate boundary was modulated by the plate-boundary dynamics and by the modes of tectonic reactivation of the intracontinental weak zones of Central Iran and at its tectonic boundaries.

KEYWORDS
exhumation, intraplate deformation, Iran, strike-slip tectonics, thermochronology

1 | INTRODUCTION

The Arabia–Eurasia collision zone is one of the most spectacular example of continental convergence on Earth (Figure 1). It consists of an amalgamation of continental blocks assembled during the long-lasting history of convergence between Eurasia and various Gondwanian continental fragments (Alavi, 1991; Bagheri & Stampfli, 2008; Berberian & King, 1981; Davoudzadeh, Soffel, & Schmidt, 1981; Ramezani & Tucker, 2003; Stocklin, 1968; Takin, 1972; Zanchi, Berra, Mattei, Ghassemi, & Sabouri, 2006). The tectonic evolution of these terrains is intimately connected with the Palaeozoic–Mesozoic evolution of the Tethyan realm along the southern Eurasia margin (e.g., Bagheri & Stampfli, 2008; Barrière et al., 2008; Mattei, Cifelli, Muttoni, & Rashid, 2015; McCall, 1997; Robert et al., 2014; Rossetti et al., 2014, 2010; Şengör, Altınır, Cin, Ustuömer, & Hsu, 1988; Stampfli & Borel, 2002). In particular, since Jurassic times, the
geodynamic setting pertained to the upper-plate domain of the Neotethyan subduction system, in the transition from oceanic subduction along the Zagros convergence zone and back-arc extension during Mesozoic times, to continental collision at the Eocene-Oligocene boundary (Agard et al., 2011; Agard, Omrani, Jolivet, & Moutheureau, 2005; Allen & Armstrong, 2008; Boulton & Robertson, 2007; Hafkenscheid, Wortei, & Spakman, 2006; Hessami, Koyi, Talbot, Tabasi, & Shabanian, 2003; Honke et al., 2009, 2010; Jolivet & Faccenna, 2000; McQuarrie, Stock, Verdel, & Wernicke, 2003; Morley et al., 2009; Moutheureau, Lacombe, & Verges, 2012; Robertson et al., 2006; Rossetti et al., 2014; Vincent et al., 2005; Vincent, Morton, Carter, Gibbs, & Baradadze, 2007).

In the Iran region, the tectonic boundaries of the continental blocks that make-up the hinterland domain of the Zagros convergence zone record a prolonged history of deformation and reactivation since Neoproterozoic times (Bagheri, Madhanifar, & Zahabi, 2016; Kargarzabanlighi, Foeken, Guest, & Stuart, 2012; Nozaem et al., 2013; Ramezani & Tucker, 2003; Rossetti et al., 2015; Tadayon et al., 2017; Verdel et al., 2007). Such tectonic zones are presently dominated by wrench tectonics that accommodates and transfers the intraplate residual convergence transmitted from the collisional boundary across the Central Iran region (the assembly of the tectonic zones known as Central Domain and the Central-East Iranian Microcontinent [hereafter referred as CEIM]; Figure 1) (Khodaverdian, Zafarani, & Rahimian, 2015; Nilforoushan et al., 2003; Vernant et al., 2004; Walpersdorf et al., 2014), in response to a geodynamic scenario dominated by −N-S-directed convergence since at least 52 Ma (Berberian, 1976, 1977; Berberian & Yeats, 1999; Jackson, Haines, & Holt, 1995; McQuarrie et al., 2003; Nowroozi & Mohajer-Ashtiy, 1985; Walker & Jackson, 2004).

Major intracontinental strike-slip fault systems commonly document long-lived polyphase tectonic histories that are normally
assumed to represent the intraplate response to spatio-temporal changes in the state of stress at plate boundaries (e.g., Calzolari, Rossetti, et al., 2016; Storti, Holdsworth, & Salvin, 2003; Van Hinsbergen et al., 2015). In particular, such a pulsed history of tectonic activity may be potentially unraveled by the associated exhumation/erosion/subsidence cycles during nucleation and development of the strike-slip deformation systems (Benowitz et al., 2011; Calzolari, Rossetti, et al., 2016; Caizolari et al., 2018; Fitzgerald, Sorkhabi, Redfield, & Stump, 1995; Gordon, Whitney, Miller, McLean, & Seaton, 2010; Roeseke, Till, Foster, & Sample, 2007; Sylvester, 1988; Woodcock & Schubert, 1994). The timing and style of deformation along such deformation systems can thus provide insights on the long-term dynamics and evolution of the deforming plate margins. On this regard, understanding the tectono-stratigraphic evolution in the upper-plate of the Neotethyan subduction during the Mesozoic-Cenozoic fragmentation and assembly of the continental blocks that make up the CEIM (Figure 1) requires a better assessment of the space-time and kinematic evolution of its major intra-continental fault systems.

In the last decades, an increasing effort has been dedicated to the understanding of the kinematic configuration and evolution of the CEIM during Neogene-Quaternary, but little information exists regarding the long-term evolution of the major intracontinental fault systems. The Doruneh Fault (DF), also referred to as Great Kavir Fault (Javadi et al., 2015; Stöcklin & Nabavi, 1973; Tchalenko, Berberian, & Behzadi, 1973; Walker & Jackson, 2004), is commonly considered the northern mechanical boundary of the CEIM (e.g., Berberian, 1985; Farbod et al., 2011; Fattahi et al., 2007; Mattei et al., 2012; Walker & Jackson, 2004; Waisersdorf et al., 2014) (Figure 1). Due to evidence indicating major kinematic shifts from dextral to sinistral shearing and changes in the regional state of stress during Late Cenozoic times, the DF region (Bagheri et al., 2016; Farbod et al., 2011; Javadi et al., 2013, 2015; Tadayon et al., 2017) represents a key region to decipher the Cenozoic spatio-temporal tectonic evolution of the CEIM. Palaeotectonic reconstructions also outline the importance of the DF as a major transfer zone in the upper-plate domain of the Neotethyan subduction (Barrier et al., 2008), which is held to be responsible for having accommodated the Mesozoic-Cenozoic rigid block rotation of the CEIM that may reconcile the post-Cimmerian distribution of the Palaeotethyan sutures in central and northeastern Iran (Alavi, Vaziri, Seyed-Emami, & Lasemi, 1997; Bagheri & Stampfli, 2008; Mattei et al., 2015; Zanchi et al., 2015). Nevertheless, uncertainty still exists regarding the role of the intraplate DF region during the re-organisation of the Arabia-Eurasia plate boundary in the transition from oceanic subduction to continental collision in Mesozoic-Cenozoic times.

This paper describes the tectonic evolution of the DF region as reconstructed from the exhumation history of pre-Cenozoic bedrock and the tectono-stratigraphic evolution of the Cenozoic and Neogene sedimentary successions exposed to the northwest of the active trace of the sinistral DF (Figures 1, 2). These results complement those presented in Tadayon et al. (2017) for the eastern region of the DF, focused on the post-Eocene history (Figure 2). When integrated with the available thermochronological data and tectonic correlations at regional scale, this allows to document a polyphase tectonic evolution during the development of a major intracontinental dextral shear belt. Lastly, the reconstructed tectonic evolution is used to link the Late Mesozoic-Cenozoic intraplate deformation of Central Iran to the tectonic re-organisation of the Zagros collisional zone.

2 GEOLOGICAL BACKGROUND

The CEIM is made of three fault-bounded N-S-oriented continental blocks (from east to west: Lut, Tabas, and Yazd), assembled in the upper-plate domain of the Neotethyan subduction zone and now forming the intra-plate domain of the Arabia-Eurasia collision zone (Alavi, 1991; Bagheri & Stampfli, 2008; Berberian & King, 1981; Davoudzadeh et al., 1981; Ramezani & Tucker, 2003; Stocklin, 1968; Takin, 1972).

The Mesozoic-Cenozoic tectonic evolution of CEIM is dominated by the polyphase opening and closure of various back-arc oceanic basins formed in the upper-plate domain of the Neotethyan subduction system (i.e., the Sabzevar, Nain Baft, Sistan oceans; Arvin & Robinson, 1994; Barrier et al., 2008; McCall, 1997; Rossetti et al., 2014; Şengör et al., 1988; Shafai Moghadam, Whitechurch, Rahgoshay, & Morsef, 2009; Shafai Moghadam, Corfu, & Stern, 2013; Stampfli & Borel, 2002; Stöcklin, 1974). Such domains presently form the various ophiolitic sutures zones that surround the CEIM (Figure 1). Available geochronological data frame the oceanic crust generation and destruction in the back-arc domain during the Cretaceous-Paleocene times (Bröcker et al., 2013; Rossetti et al., 2010, 2014; Shafai Moghadam et al., 2013; Shafai Moghadam et al., 2009, 2015; Zarrinkoub et al., 2012). Furthermore, a period of Cretaceous to Paleocene contraction is regionally documented by field evidence (Berberian & Berberian, 1981; Berberian & King, 1981; Şengör, 1990; Stocklin, 1968) and supported by tectonostratigraphical results (Kargarbanbafghi, Neubauer, & Genser, 2015; Verdel et al., 2007).

A stage of core-complex extension associated with magmatic flare-up is documented during the middle Eocene times in Central Iran (ca. 45-40 Ma; Kargarbanbafghi et al., 2012, 2015; Shafai Moghadam et al., 2014; Verdel et al., 2007; Verdel, Wernicke, Hassanzadeh, & Guest, 2011). This is followed and overprinted by renewed shortening, commonly associated with continental collision, estimated to have started at the Eocene-Oligocene boundary (e.g., Allen & Armstrong, 2008; Francois et al., 2014; Hafkenscheid et al., 2006; Hessami et al., 2001; Honke et al., 2009, 2010; Kargarbanbafghi et al., 2012; Madanipour et al., 2013; McQuarrie et al., 2003; Mouchereau et al., 2012; Robertson et al., 2005; Vincent et al., 2007). This contractional phase is well documented in the DF region, as attested by an episodic of enhanced rock exhumation to the east of the study area (Tadayon et al., 2017).

The early-middle Miocene corresponds to a period of enhanced rock exhumation and regional uplift, erosion, and deposition during the collision-related reactivation of the Kashmir-Kerman Tectonic Zone (Kargarbanbafghi et al., 2012; Verdel et al., 2007). This phase is accompanied by (a) activation of major dextral strike-slip fault zones along the northwestern boundary of the Lut Block (Calzolari, Rossetti, et al., 2016; Nozaem et al., 2013; Verdel et al., 2007) and (b) resumed tectonic activity in the DF region (Tadayon et al., 2017) (Figure 1). This event is also recorded at the regional scale, as documented along the
Bitlis-Zagros collisional zone and the Alborz Mountains (Allen et al., 2004; Axen, Lam, Grove, Stockli, & Hassanzadeh, 2001; Ballato et al., 2008, 2011, 2013; François et al., 2014; Gavillet, Axen, Stockli, Horton, & Falahari, 2010; Guest, Stockli, et al., 2006; Hessami et al., 2001; Homke et al., 2010; Khadivi, 2010; Khadivi, Moureth, Barbarand, Adatte, & Lacombe, 2012; Madanipour et al., 2013; Madanipour, Ehlers, Yassagh, & Enkelmann, 2017; Morley et al., 2009; Moureth et al., 2007; Okay, Zattin, & Cavazza, 2010), commonly referred to the transition from a juvenile to a mature stage of continental collision (hard collision in Ballato et al., 2008 and Madanipour et al., 2017).

A major tectonic reorganization occurred in Central Iran during the Late Miocene–early Pliocene as attested by (a) enhanced exhumation in the Alborz and Talesh mountains (Axen et al., 2001; Madanipour et al., 2013; Rezaei, Carter, Hovius, & Allen, 2012), (b) changes in the regional states of stress in the Kopeh Dagh region (Javidfakh, Bellier, Shabanian, Ahmadian, & Saidi, 2011; Robert et al., 2014; Shabanian, Acocella, Gioncada, Ghosimi, & Bellier, 2012; Shabanian, Bellier, Abbassi, Siame, & Farbod, 2010), (c) orocinal bending of the Alborz Mountains (Cifelli, Ballato, Alamchamadian, Sabourl, & Mattei, 2015; Mattei et al., 2017), (d) activation of the Zagros-Makran transfer zone (Regard et al., 2004) (Figure 1), and (e) the collisional lithosphere overthickening of the Zagros convergence zone that became unable to sustain further shortening (Allen et al., 2004; Austermann & Iaffaldano, 2013). This time frame is also relevant for the history of the DF, as documented by (a) fault kinematic changes within the DF region (Bagheri et al., 2016; Javadi et al., 2013, 2015; Tadayon et al., 2017) in consequence of a major shift in the regional paleoσ1 direction from NW–SE to N–S-oriented (Tadayon et al., 2017) and (b) renewed activity of dextral shearing to the south of the DF, along the Neogene–Quaternary Kuh-Sarangi-Kuh-e-Faghan fault system (Calzolari, Dela Seta, et al., 2016; Calzolari, Rossetti, et al., 2016; Calzolari et al., 2018).

The CEIM is presently moving northward at 6–13 mm/year with respect to stable Afghanistan (Walpersdorf et al., 2014) and this differential motion is largely accommodated by the active strike-slip faults systems bounding and dissecting the CEIM. These fault systems are organized into N-S dextral (from west to east: the Deshir, Anar, Nayband–Gawk, and Nehbandan faults) and E–W sinistral (from north to south: Doruneh and Dasht-e Bayaz faults) fault strands (Allen et al., 2004, 2011; Berberian, 2014; Berberian & King, 1981; Farbod et al., 2011; Fattahi et al., 2007; Masson et al., 2007; Vernant et al., 2004;
Walker & Jackson, 2004) (Figure 1). This kinematic scenario is assumed to be active since about 5 Ma (Allen et al., 2004), with the sinistral DF commonly considered to accommodate part of the shortening transmitted across the deforming CEIM (Berberian, 1976; Farbod et al., 2011, 2016; Fattahi et al., 2007; Mousavi et al., 2013; Stöcklin & Nabavi, 1973; Tchalenko et al., 1973; Walker & Jackson, 2004) (Figure 1).

Based on structural and seismological data, a tectonic scenario of dynamic rupture associated with rigid block rotations is commonly adopted to accommodate the continuous N-S convergence of Arabia-Eurasia during the Neogene-Quaternary and to frame fault zone nucleation and development during the Cenozoic evolution of the CEIM (Allen et al., 2011; Jackson et al., 1995; Jackson & McKenzie, 1984, 1988; Mattei et al., 2012; Walker & Jackson, 2004). Nonetheless, evidence of distributed pre-Pliocene dextral wrench tectonics across Central Iran (Bagheri et al., 2016; Calzolari, Della Seta, et al., 2014, Calzolari et al., 2018; Javadi et al., 2013, 2015; Nozaem et al., 2013; Tadayon et al., 2017) and the occurrence of Quaternary kinematic shifts along the DF (Farbod et al., 2011; Javadi et al., 2015) suggests that the block rotation model alone is not adequate to explain distribution and evolution of the intracontinental deformation within the CEIM. In particular, Javadi et al. (2013) documented km-scale superimposed folds produced during two phases of regional shortening, trending NW-SE (older), and NE-SW (younger), which are interpreted as connected to the shear sense switching (from dextral to sinistral) along the DF in post-Miocene times. Dextral shearing is assumed to start at the Paleocene-Eocene boundary, with the arcuate shape of the active trace of the fault resulting from the 35° counterclockwise rotation of the CEIM (Javadi et al., 2013). Through integration of structural and low-temperature thermochronological data, Tadayon et al. (2017) reconstructed a post-Eocene polyphase tectonic evolution, dominated by dextral shearing, with a major phase of enhanced cooling/exhumation starting at the Miocene/Pliocene boundary, in response to a major switch of the maximum paleo-stress direction from NW-SE to N-S.

Geodetic measurements (Mousavi et al., 2013) and solutions of earthquake focal mechanism (Shabanian et al., 2010; Zamani, Angelier, & Zamani, 2008) suggest that the N-S shortening is still active and responsible for the current stress regime and seismotectonic scenario within the DF region (see also Farbod et al., 2011). The seismotectonic scenario is characterized by fault segmenting and along-strike kinematic variation from reverse to sinistral faulting along the DF, kinematically linked to the NE-SW sinistral Dahan-Qaleh Fault and interrupted against the NNW-SSE-oriented Kharturan reverse fault (Farbod et al., 2011, 2016; Pezzo et al., 2012) (Figure 2).

3 MATERIALS AND METHODS

In this study, structural field work is integrated with apatite thermochronology from pre-Cenozoic bedrock in order to link structures to the major episodes of rock exhumation/uplift and hence to reconstruct the long-term evolution of the DF region.

Field work was aimed at refining the tectono-stratigraphic evolution of the area to the north of the central segment of DF region, by focusing on the Late Mesozoic and Cenozoic units. By integrating geological mapping and aerial photointerpretation with the available geological maps, new structural maps were produced for the study area (Figures 2, 3). The following geological maps, published by the Geological Survey of Iran, were implemented in this study: (a) the Kashmar quadrangle (at 1:250,000 scale; Eftekhar-Nezhad, Aghanabati, Hamzehpour, & Baroyant, 1976) and (b) the Doruneh (Ghaemi & Mussavi Herami, 2008), Bardaskan (Shahrahi et al., 2005), Kashmar (Taheri et al., 1998), Feyz Abad (Behroozi et al., 1987), Darin (Vahdati Daneshmand & Namad, 1999), Sheshmand (Jafarian et al., 2000), Shomkhan (Naderi Mighan et al., 1997), and Kedkan quadrangles (Naderi Mighan et al., 1998) (at 1:100,000 scale).

Stratigraphic data were collected along several across-strike transects (approximately NW-SE oriented) distributed in the study area in order to characterise the Late Mesozoic and Cenozoic sedimentary deposits. Sedimentological characteristics where documented through a series of stick-logs that record unconformities, stratigraphic relationships, and thicknesses variations through the entire Cenozoic sedimentary successions. Attribution of lithologies to specific stratigraphic units was based on the existing cartography (Behroozi et al., 1987; Ghaemi & Mussavi Herami, 2008; Shahrahi et al., 2005; Taheri et al., 1998; Vahdati Daneshmand & Namad, 1999) and correlation at regional scale (Aghanabati, 2004).

Structural investigation was devoted to define the structural architecture at the regional scale by multi-scale observations. Recognition and characterisation of the brittle structural fabrics and description of fault patterns and their kinematics was based on classical criteria as derived from the analysis of the striated fault surfaces (e.g., Doblas, 1998; Fossen, 2010; Petit, 1987), complemented by recognition of geological offsets in the field and as derived from remote sensing analysis. Structural data on faults, joints, veins, and bedding were collected at 57 geo-referenced field analysis sites distributed in the study area. A total of 543 data (including fault, fold, and bedding attitudes) were collected. Stereographic projections and analysis of structural data were performed using the software DAISY (http://host.uniroma3.it/progetti/rlab/Downloads/Programs/).

The apatite thermochronology includes apatite fission track (AFT) partial annealing zone [PAZ]: 75–120 °C, with an effective closure temperature of 110 ± 10 °C; Green & Duddy, 1989; Ketcham, Donelick, & Carlson, 1999), and apatite (U-Th)/He (AHe, partial retention zone [PRZ]: 40–80 °C, with an effective closure temperature of 70 ± 10 °C; Farley, 2000) techniques were employed to assess the cooling history in the upper 3–5 km of the crust, where brittle deformation prevails (e.g., Fitzgerald, Baldwin, Webb, & O’Sullivan, 2006; Malusà, Polino, & Zattin, 2009). The analytical procedure is described in the Appendix A.

4 STRATIGRAPHY OF THE DORUNEH FAULT REGION

Based on the existing cartography (Eftekhar-Nezhad et al., 1976; Ghaemi & Mussavi Herami, 2008; Jafarian et al., 2000; Naderi Mighan et al., 1997, 1998; Shahrahi et al., 2005; Taheri et al., 1998; Vahdati Daneshmand & Namad, 1999), the stratigraphy of the study area can be divided into two main stratigraphic groups (see the stratigraphic
column in Figure 4); (a) Precambrian to Palaeozoic basement rocks and (b) post-Cimmerian cover rocks. The latter can be further divided into (a) pre-breakup (Upper Jurassic Shemshak Fm and Lower Cretaceous conglomerates and Orbitolina-bearing limestones), (b) syn-breakup (Upper Cretaceous ophiolite-bearing successions), and (c) post-breakup Cenozoic successions.

4.1 Basement (Precambrian–Palaeozoic)

The basement rocks correspond to the polymetallic metallogenic belt of the Taknar zone (Karimpoor & Malekzadeh Shafaroudi, 2005), which is exposed into a narrow NE–SW-directed structural domain, bounded by the NE–SW Taknar Fault to the north, and the Keriz Fault to the south, respectively (Figures 2, 3). The oldest rocks in the area are exposed in an extensive outcrop of Precambrian metamorphic and magmatic rocks located to the north of the Bardaskan city (Figures 2, 3), known as Taknar Fm (Eftekhari–Nezhad et al., 1976; Shahrabi et al., 2005). The Taknar Fm consists of Late Proterozoic metasediments, crystalline limestones, green chlorite schists, tuff, sandstones, metarhyolites, and basalts, which are intruded by the Neo-Proterozoic Bonaward granitoid complex (ca. 550 Ma based on zircon U–Pb dating; Monazami Bagherzadeh et al., 2015). The Cambrian Soltaniyeh dolomites and recrystallised limestones unconformably lie on the Taknar Fm (Eftekhari–Nezhad et al., 1976; Ghaemi & Mussavi Herami, 2008; Shahrabi et al., 2005). They are followed by a discontinuous Palaeozoic, dominantly carbonate succession, spanning in age from Ordovician to Permian (Eftekhari–Nezhad et al., 1976; Shahrabi et al., 2005; Taheri et al., 1998).

4.2 Post-Cimmerian units

4.2.1 Pre-breakup units (Mesozoic)

The basement rocks are unconformably covered by Upper Jurassic shales with interlayered limestones (Ghaemi & Mussavi Herami, 2008; Shahrabi et al., 2005; Taheri et al., 1998), which crop out along a NE–SW, fault-bounded ribbon to the north of Bardaskan (Figures 2, 3). This is followed by Lower Cretaceous sediments consisting of basal red conglomerates passing into a deep marine succession made of
thin-bedded black shale, pelagic Calpionella limestones and Aiptian, bedded to massive Orbitolina-bearing limestones (Behroozi et al., 1987; Shahrbab et al., 2005). The Lower Cretaceous deposits are in fault contact both with the Upper Cretaceous ophiolite-bearing deep marine sediments, interpreted to mark the Mesozoic rifting and continental breakup in the region (Lindenberg & Jacobshagen, 1983; Maghfouri, Rastad, Mousivand, Lin, & Zaw, 2016) and the Cenozoic successions (Figures 2, 3).

4.2.2 Syn-breakup units (Upper Cretaceous)

The syn-breakup ophiolite units (known in the area as “coloured mélange” units) are assumed to be part of the Sabzevar structural zone (Figure 1). They consist of basal ultramafic suites with a volcano-sedimentary cover succession that includes cherts and Globotruncana-bearing carbonates, stratigraphically attributed to the Upper Cretaceous (Eftekhari-Nezhad et al., 1976; Ghaemi & Mussavi Herami, 2008; Maghfouri et al., 2016; Shahrbab et al., 2005). Various gabbroic subvolcanic (sill-like) intrusions occur within the basal volcano-sedimentary unit. The geochemical fingerprint of the effusive and intrusive magmatic suite suggests a back-arc palaeotectonic environment of formation during the Cretaceous transition from continental rifting to continental breakup in the upper-plaue of the Neotethyan subduction (Maghfouri et al., 2016).

The ophiolite units are mostly exposed within a NE-SW trending structural high, bounded by two major sub-vertical faults, the Taknar (Rivash) Fault to the south and the Sebeh Fault to the north, respectively (Figures 2, 3). To the southeast, the NW-dipping Taknar Fault bounds at high-angle the contact with the basement units of the Taknar zone. To the northwest, the SE-dipping Sebeh Fault controls the contact between the ophiolite units and the Cenozoic volcano-sedimentary deposits (see below).

4.2.3 Post-breakup units (Cenozoic)

The basement and Mesozoic cover rocks units are unconformably covered by a thick pile of Cenozoic deposits. The stratigraphic relations as reconstructed in the field are shown in Figures 4 and 5. Within the Cenozoic sedimentary pile, six main regional unconformities were recognised in the field (numbered as 1–6 in Figure 4).

The Cenozoic succession starts with an up 250-m thick pile of crudely stratified red conglomerates passing to grey conglomerates and sandstones with sporadic thin interbedded layers of Lower-Middle Palaeocene Missetania sp. bearing limestone that lie unconformably onto the coloured mélange rocks (Ghaemi & Mussavi Herami, 2008; Unconformity 1 in Figure 5a). The clasts, making up the bulk of the basal deposits, are derived from the proximal ophiolite-bearing units, as documented by the occurrence of angular to rounded (up to approximately 70 cm in diameter) cobbles made of gabbror, basalt, and Globotruncana limestones (Figure 5a). These deposits mainly crop out to the north of the Sebeh Fault, in fault contact with ophiolite-bearing units (Figure 5b). They can be stratigraphically correlated to the Kerman Fm (Aghanabadi, 2004).

A thick pile of Upper Palaeocene-Eocene flysch-type volcanoclastic deposits and volcanic products, passing upward to nummulitic limestones and green shales and marls (Shahrbab et al., 2005; Taheri et al., 1998) covers unconformably the Kerman Fm (unconformity 2 in Figure 5a). This Palaeocene-Eocene succession is
dominantly exposed in the NW sector of the study area, where it shows a maximum thickness of approximately 5 km (Figure 2). To the east, it is intruded by the middle Eocene calcalkaline Kashmir granitoids (Shafaii Moghadam et al., 2015; Sotani, 2000; Taheri et al., 1998), which define a approximately 100-km long magmatic belt (hereafter referred to as Kashmir–Azghand Intrusive complex, KAIC) to the north of the Kashmir and Azghand cities (Figure 2). The U-Pb zircon dating of the southern rim of the KAIC provided a middle Eocene (40–41 Ma) formation age for the Kashmir granitoids (Shafaii Moghadam et al., 2015). It is worth noting that the Palaeocene–Eocene strata also unconformably cover the folded and sheared ophiolite units (Figures 5a-b). Vaehadi Daneshmand & Nadim, 1999) (Figure 3) and contain abundant clasts of Palaeocene–Eocene volcanic rocks (Figure 5b). To the east, despite the original contacts are reworked by faulting, a major erosional unconformity can be recognized between the KAIC and the basal deposits of the Lower Red Fm, which also contain boulders (up to 1 m in diameter) of the Eocene KAIC granitoids. This field evidence indicates that the KAIC was already eroded and exposed to erosion prior to the sedimentation of the Lower Red Fm (Tadayon et al., 2017).

The Late Oligocene–early Miocene marine deposits of the Qom fm (QF; Amirshahkarami & Karavan, 2015; Reuter et al., 2007), made of thin-beded fossiliferous marine limestone, light green marl with interlayers of tufaceous sandstone, and gyposiferous horizons, are discontinuously exposed in the north-western sector of the study area (Figure 3; Unconformity 4 in Figure 4). The Qom Fm unconformably lies on top of the Palaeocene–Eocene strata in Mohammad Zurub and to the north of the Doruneh village (Ghaemi & Mussavi Herami, 2008), whereas the observed contacts with the LRF are characterised by a low-angle angular unconformity (Figure 5d).
The Miocene continental deposits of the Upper Red Fm (URF) unconformably cover the QF in Chah Talkh to the NW of the study area (Figure 3; Unconformity 5 in Figure 5e). In the southern sector of the study area, these deposits are discontinuous and are systematically observed in tectonic contact along faults with the previously described units (Behrooz et al., 1987; Farbod et al., 2011, 2016; Ghaemi & Mussavi Herami, 2008; Vahdati Daneshmand & Nadim, 1999) (Figures 2 and 3).

Pliocene–Quaternary continental deposits consisting of grey unconsolidated conglomerate and gravels cover the Upper Red Fm above a major erosional and angular unconformity (Unconformity 6 in Figure 5f). Quaternary deposits are affected by diffuse faulting and deformation (Farbod et al., 2011, 2016; Fattahi et al., 2007; Tadayon et al., 2017; Taheri et al., 1998).

5 | STRUCTURAL ARCHITECTURE

The Figure 2 shows the active trace of the sinistral DF and the region to the north of its central portion that is dominated by a huge NE–SW trending deformation zone, in part described in Javadi et al. (2013). The deformation zone is made of an about 40-km thick, verticalized panel of rock successions, where the pre-Cenozoic rocks are delimited by sub-vertical faults: the Keriz Fault, to the south and the Sebeh and Kalat-e-Bargh faults, to the north. The internal structure is characterized by various fault-bounded units arranged to form a plurikilometric-scale strike-slip duplex structure at the map scale (Woodcock & Fischer, 1986) that involves the basement, ophiolite, and Cenozoic units (Figure 3). A NE–SW-trending, fault-bounded structural high forms the backbone of the study area. Here, the Upper Cretaceous ophiolite units are in fault contact to the NW with the Cenozoic units through the Kalat-e-Bargh-Sebeh Fault and to the SE with the basement units and their Mesozoic cover rocks of the Takanar structural zone through the Taknar Fault (Shahrbaf et al., 2005), respectively (Figure 3).

The along-strike continuity of the deformation zone is interrupted by the sub-vertical, NNE–SSW striking, active sinistral Dahan-Qaleh Fault (Farbod et al., 2011, 2016). The deformation zone thins towards the southwest, where the thickness of the deformation zone is about 6 km. It terminates against the NNW–SSE striking, active reverse Kharturan Fault, which is assumed to absorb the southwest motion of northern block of the DF to the west of the Dahan-Qaleh Fault (Farbod et al., 2011; Pezzo et al., 2012). The deformation zone fades out toward the east, cutting across the middle Eocene KAIC (Tadayon et al., 2017).

To illustrate the overall structural architecture, fault kinematics, and relative age of faulting and folding, in the following sections, we describe three representative geological cross sections traced perpendicular to the main structures documented in the field (A–A’, B–B’ and C–C’ in Figure 3). In particular, cross-section A–A’ and B–B’ (Figures 6–8) illustrate the structures across the structural high and cross-section C–C’ illustrates the style of folding that affects the Cenozoic units exposed in the northwest of the study area (Figure 9).

5.1 | Cross-section A–A’

The ophiolite-bearing structural high is delimited by two major sub-vertical faults, the SE-dipping Kalat-e-Bargh Fault and the NW-dipping Taknar Fault, located to the north and to the south, respectively (Figure 6a). The overall fault arrangement and structures describes sub-vertical fault noses, geometrically resembling those of positive flower structures (see e.g., Sylvester, 1988).

The Kalat-e-Bargh Fault juxtaposes the Upper Cretaceous sedimentary cover rocks of the ophiolite units in the hangingwall onto the Eocene volcanoclastic deposits in the footwall. The fault contact is marked by approximately 100-m thick fault damage zone that can be followed along strike for more than 50 km (Figure 3). A prominent, up to 40-m thick, horizon of S-C tectonites formed at the expenses of Globotruncana limestones is observed in the immediate hangingwall of the Kalat-e-Bargh Fault (Figures 6a and 7a–c). Fault kinematics as deduced by S-C tectonites, growth fibres and synthetic Riedel shears, points to a dominant dip-slip movement with a minor dextral component (average sickenline pitch values of 72°) and an overall top-to-

![Figure 6](https://example.com/figure6.png)  
**Figure 6** Interpretative geological cross-section across the NE–SW structural high bounded by the Sebeh-Kalat-e-Bargh and Taknar faults (see Figure 3 for location of the geological cross-section). (a) Trace AA’. (b) Trace BB’. The location of figures showing representative structures in the field are also indicated [Colour figure can be viewed at wileyonlinelibrary.com]
FIGURE 7  Structural evidence across the geological cross-section A–A’. (a) Satellite image (GeoEye data base) showing the structural architecture of the dextral transpressional NE–SW Kalat-e-Barg Fault, which control the contact between the Cretaceous ophiolite rocks and the folded Eocene deposits. Note the NE–SW drag syncline at the fault footwall. The stereoplot (Schmidt net, lower hemisphere projection) shows the corresponding structural data. The yellow rectangle shows location of Figure 7b. (b) The Kalat-e-Barg Fault in the field. The fault core is decorated by S-C tectonites indicating top to the NW sense of shear (Figure 7c). (d) Panoramic view looking NE showing the damage zone of the Taknar Fault, which corresponds to the reactivated steeply-dipping fault contact between the Lower Cretaceous limestones (fault footwall) and the Upper Cretaceous ophiolite units (fault hangingwall). The stereoplot (Schmidt net, lower hemisphere projection) shows the corresponding structural data, which attests for dominant reverse kinematics. (e) Satellite image (GeoEye data base; see Figure 2a and 6a for location) showing a set of en-echelon, NE trending folds affecting the Lower Red Fm (LRF) deposits that are cut across by ENE–WSW dextral transpressive faults. The lithons of Jurassic and Miocene URF deposits occur along the northern boundary fault. NE–SW dextral and NW–SE sinistral conjugate fault cut across the ENE–WSW fault zone (see Figure 7d). (f–g) The NW-dipping dextral transpressive fault contact that controls the contact between the Cambrian massive dolomites of the Solanieh Fm (continental basement) with the Oligocene LRF (see Figure 7e for location). The stereoplot (Schmidt net, lower hemisphere projection) shows the corresponding structural data. (h) The fault damage zone that controls the sub-vertical contact between the LRF (conglomerate and sandstones) and the URF (green marls). Camera shows the location and view direction of Figure 7i. (i) The fault core of the LRF–URF contact is decorated by approximately 1–m thick sub-vertical panel of S-C tectonites, developed at the expenses of the URF marly deposits. The synthetic Riedel shears and the slickenlines on a polished fault surface are also indicated. The stereoplot (Schmidt net, lower hemisphere projection) shows the corresponding structural data, attesting for dextral shearing. (j) Conjugate set of NW–SE dextral and NNE–SSW sinistral fault strands which cut across the LRF sandstones (plane view). The acute dihedral angle defined by these conjugate fault strands are bisected by roughly N–S striking vein arrays. The stereoplot (Schmidt net, lower hemisphere projection) shows the corresponding structural data (for location see Figure 7e) [Colour figure can be viewed at wileyonlinelibrary.com]

the NW sense of tectonic transport (Figures 7a,b). In the footwall, the Eocene volcanoclastic deposits are deformed to form an array of ENE–SSW trending folds, cut by the Kalat-e-Barg Fault (Figures 6a and 7a). A drag syncline lies immediately below the footwall area of the Kalat-e-Barg Fault (Figures 3 and 7a).

The Taknar Fault is defined by a more than 100-m thick, NW–SE damage zone that can be almost continuously followed in the field for more than 100 km. The fault zone is characterized by high angle the contact between the ophiolite units, at its hangingwall, and the various stratigraphic units that make up the Taknar zone, at its footwall. As defined at many localities, the fault kinematics along the main trace of the Taknar Fault is oblique dextral and/or dip-slip reverse (slickenline pitch angle ranging 53°–72°), attesting for an overall top-to-the-ESE sense of tectonic transport (Figure 7d).

Moving toward the southeast, units located in the footwall of the Taknar Fault are bounded by NW-dipping fault splays, which are interpreted to converge at depth into the main bounding fault (Figure 6a). The southeastern front of the strike-slip duplex structure at the footwall of the Taknar Fault is characterized by a set of sub-vertical, NS0°–70°-striking NW-dipping fault systems. Moving from
NW to SE, these fault strands control the contact between the Cambrian massive dolomites of the Soltanieh Fm with the siliciclastic deposits of the Oligocene LRF and between these deposits and the marly deposits of the Miocene URF (Figures 6a and 7e–g). Notably, the deposits of the LRF cropping out in the intrafault block domain are deformed into a set of en-echelon NE trending folds that are abruptly cut across by major N70°-87° striking fault systems. The kinematics of these fault systems was systematically investigated in the field. Polished fault surfaces show oblique slickenlines (slickenline pitch angle ranging 50°-75°) and the kinematic criteria as provided by fault drags, growth fibres, and synthetic Riedel shears are consistent with dextral reverse sense of motion (top-to-the ESE). Particularly significant is the fault architecture bounding the LRF/URF contact towards the Quaternary plain. The fault contact is characterised by a 30-m thick, N75°-oriented sub-vertical damage zone, with an approximately 10-m thick fault core, mostly developed at the expenses of the marly deposits of the Upper Red Fm. The fault core is decorated by approximately 1-m thick sub-vertical panel of S-C tectonites. Slickenlines on the C and polished fault surfaces are dominantly sub-horizontal (pitch angle ranging 3°-15°); the sense of shear is dextral (Figure 7e and 7h–i). This early generation of dextral/reverse faulting is overprinted by a subsequent generation of faults, consisting of a conjugate set of NNW-SSE dextral and NNE-SSW sinistral fault strands. At the meso-scale, the dihedral angle between these faults is bisected by roughly N-S striking vein arrays (Figure 7j). Cross-cutting relationships between the two fault generations are evident at regional scale and in the field and confirm the relative age of the two faulting events (Figures 7e and 7k).

5.2 | Cross-section B-B’

The structural architecture across cross-section B-B’ is dominated by a major antiformal structure formed by Upper Cretaceous ophiolite
units. They occur within a push-up horst (Woodcock & Rickards, 2003) in the hangingwall of the Sebeh and the Taknar Fault systems, to the NW and to the SE, respectively (Figure 6b). This structural high is unconformably covered by the Palaeocene-Eocene volcanoclastic successions (Figures 3 and 8a; Unconformity 2 in Figure 4).

The Sebeh Fault dips steeply to the SE and shows along-strike bending from N40° to N70°. It is defined by approximately 100-m thick sub-vertical parsel of brecciated fault rocks, marking-up the contact between the ophiolite units and the Eocene volcanoclastic deposits at its footwall. The fault contact is discontinuously decorated by dm-thick S-C tectonites at the fault hangingwall developed at the expenses of the Upper Cretaceous Globotruncanica limestones. The S-C fabrics systematically point to dextral reverse sense of shear (sickenline pitch angle of 128°-132°) (Figures 8bc).

The Taknar Fault strikes N70°-90°, dips steeply to the NNW and shows up to 150-m thick fault damage zone (Figures 8d,e). In map view, the trace of the Taknar Fault is defined by strike-slip duplexes, mostly developed at the expenses of the Lower Cretaceous Orbitolina limestones (Figure 3). The main contact is defined by sub-vertical fault surfaces, generally dipping NNW and showing two generations of slickenlines: an early dip-parallel (sickenline pitch angle: 102°−135°) generation, showing reverse/dextral (top-to-the-SE) sense of shear, overprinted by a sub-horizontal (sickenline pitch angle: 0°−175°) one, showing dextral kinematics (Figure 8f).

The intrafault block exposed within the erosional window below the Palaeocene-Eocene cover is affected by minor NE-SW faults, which crosscut folded and tilted bedding (dipping steeply either to the NW or to the SE) of the ophiolite units (Figure 8a). These units are intruded by magmatic rocks, consisting of steeply-dipping, layer-parallel (sill-like) gabbroic to granodioritic intrusive sheets (from ~50 to 1,000 m in thickness) that are considered part of the original ophiolite sequence (Maghfouri et al., 2016). These intrusive bodies...
together with the host rocks are tightly folded to form km-sized, NE-SW trending chevron-type folds, which produce a repetition of the original lithological sequence. These folds (hereafter referred as F1) trend oblique to the main trace of the Taknar and Sebeh faults and occur within dextral shear zones to form a major strike-slip duplex (Woodcock & Fischer, 1986) (Figure 3). Contour diagram of pole to bedding for the F1 folds plotted on a π-diagram provides a π-axis trending N41°, in good accordance with the fold hinges measured in the field (Figure 8a).

A set of roughly E-W oriented, sub-vertical fault strands cut across the above described structures. In particular, these faults are observed to cut across the intrafault block bounded by the Kalat-e-Bargh and Taknar faults to the north of the Donuneh Village (Figure 3). The Taknar Fault is also dextrally segmented by an en-echelon array of E-W faults, cut across the Precambrian basement and Eocene rocks at the fault footwall (Figure 8d). The associated fault damage zones are up to 30-m thick and made of sub-vertical fault surfaces and fracture arrays. Polished fault surfaces systematically document almost pure strike-slip motion (slickenline pitch angle of 0°-10° and 170°-180°; see the stereoplot in Figure 8h) and kinematic indicators, as provided by synthetic Riedel shears and growth fibres in carbonate rocks, attest for dextral shear (Figure 8h).

5.3 Cross-section C-C′

The cross-section C-C′ runs across the Cenozoic deposits and the superimposed folds (Javadi et al., 2013) exposed at the northwestern sector of the study area (Figure 3). Based on the overprinting relationships at map scale and systematic bedding attitude survey along transects trending sub-parallel to the cross-section C-C′, two main, km-sized, post-F1 fold generations are recognised (Figure 9a): (a) an early generation (F2), with fold axes trending NE-SW and affecting the Late Palaeeocene-Eocene successions, the LRF and QF deposits and (b) a second generation (F3), with fold axes trending around E-W and recorded in the URF deposits. A late generation (F4), with fold axes trending NW-SE, is locally observed to affect only the URF (see also Javadi et al., 2013). Countouring of the poles to bedding plotted on a π-diagram provides a π-axis trending N62° (LRF and QF) and N243° (Eocene-Paleocene successions) for F2 and N80° for F3, respectively (Figure 9b).

In the pre-URF deposits, a marked difference in the F2 fold styles is reconstructed, as defined by the fold tightness, with open to close folding in the Late Palaeeocene-Eocene strata (interlimb angle: 70°-30°) and tight folding (interlimb angle: 20°-30°) in the post-Eocene LRF and QF deposits (see also Figure 3). This is well documented by the π-diagrams that show a markedly different distribution of the fold limbs, attesting for tighter fold shapes in the LRF and QF than in the Late Palaeeocene-Eocene successions (Figure 9b).

An interpretative reconstruction of the structure across section C-C′ is shown in Figure 9c. The different style of folding during F2 is interpreted as induced by detachment folding of the post-Eocene successions above a stratoagraphically fixed detachment horizon localised in the marly shales at the top of the Eocene succession. The reconstructed fan-shaped attitude of the F2 axial surfaces in the post-Eocene successions is consistent with superimposed folding at a larger wavelength during F3, as also outlined by the F2 open folding of the unconformity at the base of the URF (Figure 9c).

6 THERMOCRONOLOGY

Based on the structural and stratigraphic field investigations, six samples were collected for AFT and AHe thermochronology (Figure 3 and Tables 1 and 2). Five samples (113-08, 113-10, 114-23, 114-34, and 114-38) are from the Upper Cretaceous mafic to intermediate intrusive sills emplaced within the ophiolite units that make up the fault-bounded NE-SW structural high, at the fault hangingwall of the Taknar and Sebeh faults (Figure 3). One additional sample (F2-02) is from the Neoproterozoic Bonaward granodiorite at the footwall of the Taknar Fault, with a zircon U-Pb crystallization age of ca. 540 Ma (Monazami Baghzadeh et al., 2015) (Figure 3).

6.1 Apatite fission track

Four of the six analysed samples have the physical and chemical (e.g., U content) characteristics required to obtain reliable thermochronological AFT data. To assess the homogeneity of age population, chi-squared test has been applied on data from single samples. All of them passed the chi-squared test at the 95% confidence level, thus showing a limited single crystal age spread. Results of the AFT analysis are presented in Table 1.

Since the number of apatite crystals was low, the standard deviation of AFT ages is high, with AFT ages spanning from 93.6 ± 27.2 to 60 ± 24.5 Ma. Constrained track length measurements were not obtained for these samples, due to the low number of apatite grains and/or low U concentration (see e.g., Gleadow, Duddy, Green, & Lovering, 1986; Gleadow & Seiler, 2014). A poor correlation exists between the AFT

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Latitude/Longitude</th>
<th>Elevation (m)</th>
<th>Number of crystals</th>
<th>Spontaneous track density (×10⁵ cm⁻²)</th>
<th>Induced track density (×10⁵ cm⁻²)</th>
<th>AFT age ± σ (Ma)</th>
<th>Chi-squared test (χ²)</th>
<th>P (χ²)%</th>
</tr>
</thead>
<tbody>
<tr>
<td>114-23</td>
<td>35°47′39.35″N &lt;br&gt; E35°24′19.59″E</td>
<td>1,480</td>
<td>7</td>
<td>10</td>
<td>27</td>
<td>69.0 ± 25.8</td>
<td>1.2</td>
<td>97.6</td>
</tr>
<tr>
<td>13-08</td>
<td>35°52′31.99″N &lt;br&gt; E35°26′22.17″E</td>
<td>1,666</td>
<td>8</td>
<td>8</td>
<td>25</td>
<td>60.0 ± 24.5</td>
<td>0.1</td>
<td>100.0</td>
</tr>
<tr>
<td>113-10</td>
<td>35°52′38.34″N &lt;br&gt; E35°25′38.94″E</td>
<td>1,763</td>
<td>9</td>
<td>18</td>
<td>36</td>
<td>93.6 ± 27.2</td>
<td>2.2</td>
<td>97.3</td>
</tr>
<tr>
<td>114-34</td>
<td>35°52′31.42″N &lt;br&gt; E35°27′38.16″E</td>
<td>1,810</td>
<td>17</td>
<td>23</td>
<td>60</td>
<td>71.9 ± 17.8</td>
<td>0.8</td>
<td>100.0</td>
</tr>
</tbody>
</table>

*All calculations were computed with a Zeta number of 362.48 ± 12.69, dosimeter density (ρ-d): 1.040⁵ and number of track counted for dosimeter (N-d): 5,000.
TABLE 2  List of the studied samples for AHe thermochronology, with geographical location and analytical data for single apatite grains indicated

<table>
<thead>
<tr>
<th>Sample number</th>
<th>Sample number</th>
<th>Location: Latitude/ Longitude</th>
<th>Elevation (m)</th>
<th>Radius (µm)</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>4He (nmol/g)</th>
<th>eU (ppm)</th>
<th>Corrected single crystal AHe age (Ma) ± σ</th>
<th>mean AHe age (Ma) ± σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>F2-02-1</td>
<td>F2-02-2</td>
<td>57°50′19.17″ E-35°23′33.39″N</td>
<td>1,375</td>
<td>44.88</td>
<td>36.62</td>
<td>34.28</td>
<td>1.21</td>
<td>44.68</td>
<td>728 ± 0.51</td>
<td>5.93 ± 0.21</td>
</tr>
<tr>
<td>F2-02-3</td>
<td></td>
<td></td>
<td>49.56</td>
<td>18.83</td>
<td>27.59</td>
<td>1.38</td>
<td>25.22</td>
<td>19.77 ± 0.77</td>
<td></td>
<td></td>
</tr>
<tr>
<td>I14-23</td>
<td>I14-23-2</td>
<td>57°47′39.35″ E-35°24′15.59″N</td>
<td>1,480</td>
<td>39.00</td>
<td>2.85</td>
<td>6.59</td>
<td>0.22</td>
<td>4.40</td>
<td>14.07 ± 1.4</td>
<td>13 ± 1.9</td>
</tr>
<tr>
<td>I14-23-3</td>
<td></td>
<td></td>
<td>40.62</td>
<td>1.88</td>
<td>3.49</td>
<td>0.12</td>
<td>2.70</td>
<td>11.94 ± 2.56</td>
<td></td>
<td></td>
</tr>
<tr>
<td>I13-10</td>
<td>I13-10-3</td>
<td>57°52′38.34″ E-35°26′36.94″N</td>
<td>1,763</td>
<td>3.499</td>
<td>19.14</td>
<td>75.34</td>
<td>3.43</td>
<td>36.64</td>
<td>29 ± 2.14</td>
<td>29 ± 2.14</td>
</tr>
<tr>
<td>I14-34</td>
<td>I14-34-3</td>
<td>57°52′31.42″ E-35°27′36.16″N</td>
<td>1,810</td>
<td>39.25</td>
<td>2.52</td>
<td>5.19</td>
<td>0.29</td>
<td>3.74</td>
<td>21.57 ± 2.37</td>
<td>21.57 ± 2.37</td>
</tr>
<tr>
<td>I14-38</td>
<td>I14-38-2</td>
<td>57°52′55.82″ E-35°26′43.72″N</td>
<td>1,845</td>
<td>35.54</td>
<td>2.19</td>
<td>4.84</td>
<td>0.23</td>
<td>3.33</td>
<td>20.98 ± 3.58</td>
<td>22.87 ± 2.5</td>
</tr>
<tr>
<td>I14-38-3</td>
<td></td>
<td></td>
<td>45.43</td>
<td>2.38</td>
<td>6.02</td>
<td>0.34</td>
<td>3.79</td>
<td>23.17 ± 1.42</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

| 113-23 and 113-34 show evidence of partial resetting of the AFT system in post-Palaeocene times, yielding minimum ages younger than the Cretaceous–Palaeocene unconformity. Since the altitude of sample 113-10 is ~50 m lower than that of sample I14-34 (Figure 10) and due to the limited number of apatite grains, we can assume that sample I13-10 is partially reset too. |

6.2 | Apatite (Uranium–Thorium)/Helium thermochronology

Results of the AHe analysis, including single crystal corrected age, mean age and standard deviation, radius, U, Th, He, and eU content are listed in Table 2. The quality of the AHe results has been checked with standard criteria such as He and U concentrations, the Th/U ratio, presence of inclusions and coating, age dispersion, and comparison with AFT ages. The mean AHe ages from the hangingwall of the Taknar Fault range from the early Oligocene to the middle Miocene (mean ages spanning 13 ± 1.90 to 29 ± 1.84 Ma), whereas the youngest age is obtained at the fault footwall (sample F2-02) that provided a Late Miocene age of 5.93 ± 0.21 Ma (Figure 3 and Table 2). The AHe mean age versus elevation graph at the hangingwall of the Taknar Fault shows a general loose positive correlation, suggesting that the different samples experienced a comparable cooling/exhumation history in the last ca. 30 Ma (Figure 10).

6.3 | Time–Temperature modelling

In order to reconstruct the cooling history (T–t) of the ophiolite units within the NE–SW push-up horst (hangingwall of the Taknar and Sebeh faults; Figure 3), the AFT and AHe data were modelled by means of the software HeFTy (Ketcham, 2005). Due to the lack of confined track information, the modelling of fission-track data is based only on ages. HeFTy software uses the Monte Carlo method to find the best T–t paths that are compatible with AFT and AHe data (Ketcham, 2005; Ketcham, Donelick, Balestri, & Zattin, 2009). Thermal modelling was applied to the samples I13-10, I14-23 and I14-34, for which both AFT and AHe are available. The following constraints
**FIGURE 11** Radial plots for the AFT samples with Upper Cretaceous formation ages (85–100 Ma; green areas); the unconformity (ca. 60–65 Ma) at the base of the Palaeocene–Eocene deposits (red areas) is also indicated. Central age, relative error, and number of grains are provided in each plot. Radial plots were constructed using the TRACKKEY 4.2 software (Dunkl, 2002) [Colour figure can be viewed at wileyonlinelibrary.com]

**FIGURE 12** The T-t history of the Taknar Fault hangingwall (structural high) as provided by the thermal modelling performed with the HeFTy software (Ketcham, 2005) (see text for further details). The boxes indicate the T-t constraints adopted in this study as follows: (a) the crystallisation/formation age of the intrusive sills set at 85–100 Ma (Box-1); (b) the unconformity at the base of the Palaeocene–Eocene deposits, placed in the 65–60 Ma time lapse (Box-2); and (c) the age of the AFT resetting event is placed in the time interval 50–34 Ma (Box-3). The blue, green, and red lines refer to the weighed-mean thermal history path for three samples I13–10, I14–23 and I14–34, respectively. The green and purple points in the boxes refer to the acceptable and good constraints, respectively, used by the HeFTy software to model the T-t paths. Cooling rates have been estimated from the average slope of the T-t paths within the AFT partial annealing (120–75 °C) and AHe partial retention (80–40 °C) temperature windows [Colour figure can be viewed at wileyonlinelibrary.com]
are adopted here (Figure 12): (a) the crystallization/formation age of the analysed intrusive sills is placed in the Upper Cretaceous (100-85 Ma), based on the stratigraphic evidence presented in Maghfouri et al. (2016), (Box-1); (b) the Cretaceous-Palaeocene unconformity is placed in the 66-58 Ma time lapse (Box-2); (c) the age of the AFT resetting event is placed in the time interval 50-34 Ma (Box-3), as supported by the presence of a major erosional unconformity at the top of the Palaeocene-Eocene successions (Figure 5b); and (d) present-day mean temperature is assumed to be 15 ± 5 °C.

The thermal modelling results document a punctuated history of exhumation/burial of the hangingwall of the Taknar-Sebeh Fault. The first stage corresponds to an early episode of cooling (~4 °C/Ma) during which the intrusive bodies and the hosting ophiolite units have been exhumed to the surface during Upper Cretaceous-Palaeocene times. The second stage corresponds to a burial phase when the ophiolite units as a whole entered the PAZ in consequence of the deposition of the Palaeocene-Eocene volcaniclastic successions. The third stage corresponds to a final (nearly steady state) exhumation-related cooling (~3 °C/Ma) of the ophiolite units, started during the middle-Late Eocene (Figure 12).

7 | DISCUSSION

The structural setting and the thermochronological data sets document that the study area experienced a long-lasting and polyphase history of tectonic reactivation, following an episode of continental rifting and breakup in the Cretaceous time, linked to the opening of the Sabzevar oceanic realm (Maghfouri et al., 2016).

The regional structural and stratigraphic architecture, coupled with the relative chronology of folding (F₁, F₂, and F₃) and overprinting relationships between different fault sets, allows to reconstruct the tectonic regimes, responsible for reactivation, in space and time in the region. In particular, assuming (a) fold axis orientations as parallel to the maximum instantaneous stretching direction in the horizontal plane and (b) irrotational incremental strain (Fossen, Tikoff, & Teysier, 1994), the deformation regimes associated with the polyphase reactivation of the basin boundary fault, that is, the Sebeh and Taknar faults (Figure 3), can be reconstructed in space and time.

The early phase of basin inversion and reactivation of the early NE-SW Mesozoic rift structures developed in response to an early NW-SE directed maximum compression direction, active since Upper Cretaceous-Palaeocene times (90-60 Ma). This tectonic regime was responsible for the positive inversion of the rift shoulders with the formation of the push-up horst bounded by the Taknar and Sebeh fault systems. Within the intrafault block, shortening was accommodated through F₁ folding, tilting of the bedding, and overthrusting of the Cretaceous ophiolite units, accompanied by topographic growth, erosion, and exhumation. In the intrafault block, the angle θ (measured dockwise) between the F₁ fold trend and the Taknar Fault strike is 32°, pointing to a pure shear-dominated transpression (Fossen et al., 1994) (Figures 13a,b). The pure-versus simple-shear (convergence vs. strike-slip, sensu Krantz, 1995) component within the intrafault block can be estimated as approximately 30% and 70%, respectively, compatible with dextral transpression (Figures 13c). Based on the cooling rate calculated for the Upper Cretaceous-Palaeocene exhumation event (~4 °C/Ma; Figure 12) and assuming (a) an average paleo-geothermal gradient of 25 ± 5 °C (e.g.,

FIGURE 13 (a) Angular relationships between the mean orientation of the Taknar Fault and the hinge trends of the main three (F₁, F₂, and F₃) fold generations. (b) Angle θ between the maximum horizontal instantaneous stretching axis (ISA) and the deformation zone boundary (here, assumed as the strike of the Taknar Fault) as a function of the kinematic vorticity number Wk (modified after Fossen et al. (1994)) for the three generations of folds (symbols as in Figure 13a). (c) Fault strike versus fold hinge trend graph showing the corresponding deformation regimes for each sets of fold generation (symbols as in Figure 13a): Tp: transpression, dx: dextral, sx: sinistral. (d) Correlation diagram of vertical versus strike-slip component rate of fault slip contoured for different simple shear % (symbols as in Figure 13a) [Colour figure can be viewed at wileyonlinelibrary.com]
Kargarbafgh et al., 2012) and (b) the pure shear component of deformation as transferred to vertical exhumation, an exhumation rate of 0.16 ± 0.03 km/Ma can be estimated. Using the previously estimated convergence versus strike-slip component (as 30 and 70%, respectively) and assuming no surface uplift, a conservative long-term minimum strike-slip rate for the Taknar Fault during the Upper Cretaceous-Paleocene event can be estimated to 0.4 ± 0.1 km/Ma (Figure 13d).

The Cretaceous-Paleocene exhumation is followed by a burial phase, followed by renewed exhumation with cooling through the AHe PRZ completed during the Miocene (Figure 12). The fold architecture, contact geometries and characteristics between the LRF and the QF suggest syn-tectonic deposition during continuous NW-SE shortening and F2 folding, postdating deposition of the Palaeocene-Eocene volcaniclastic deposits. Assuming the F2 folds as structurally linked to the post-Paleocene activity of the Taknar Fault, the angle is 13°, again attesting for a pure shear-dominated regime of dextral transpression (Figures 13a-c). The pure-versus simple-shear component is estimated as 70 and 30%, respectively (Figure 13b). Based on the cooling rate (3 °C/Ma; Figure 12), an exhumation rate of 0.12 ± 0.02 km/Ma (again, assuming an average paleo-geothermal gradient of 25 ± 5 °C) can be estimated for the post-Eocene evolution. This exhumation stage corresponded to the time lapse when the push-up horst bounded by the Taknar and Sebeh faults systems passed through the AHe PRZ (Figure 12). Using the above estimated convergence and strike-slip component of the associated shear deformation and assuming no surface uplift, a conservative long-term strike-slip rate for the Taknar Fault during the Eocene-Miocene can be estimated to 0.05 ± 0.01 km/Ma (Figure 13d).

As attested by the overprinting relationships as reconstructed in the field and by satellite images (Figures 2 and 8) and similarly to what reconstructed from the central segment of the DF region (Tadayon et al., 2017), this early transpressional regime was overprinted by dextral E-W wrenching. The transition from transpressional- to wrench-dominated dextral shearing is here interpreted as the consequence of shortcut faults formation (e.g., Dooley & Schreurs, 2012) cutting across the NE-SW push-up horst bounded by the Sebeh-Kait-e-Bargh and Taknar faults. These faults formed along pre-existing structural anisotropies, oriented transverse to propagation direction of regional E-W dextral shearing, under a stress regime dominated by a continuous NW-SE-oriented maximum compression direction.

The post-Miocene tectonic evolution is dominated by a renewed shortening event, under a general N-S-oriented maximum compression direction and accommodated by E-W-oriented F2 folding, as dominantly recorded by deformation of the URF. This renewed shortening is recorded by the cooling/exhumation history at the footwall of the Taknar Fault (sample F2-02; Figure 3). Here, Late Miocene or younger AHe ages are obtained, suggesting that the post-Miocene exhumation was tectonically controlled by the new N-S-oriented shortening regime. This also suggests that the Cenozoic sedimentation in the structurally low regions is responsible for the resetting of the AHe system at the fault footwall. A N-S-directed maximum compression direction is compatible with the occurrence of conjugate NW-SE dextral and NNE-SSW sinistral strike-slip fault systems that dissect and cut across the older ENE-WSW-oriented dextral fault strands (Figure 7). This is in agreement with the evidence presented in Tadayon et al. (2017) from the eastern DF region (Figure 2), where a set of conjugate dextral NW-SE and sinistral NNE-SSW faults overprints pristine E-W dextral faults during a stage of accelerated exhumation active at the Miocene-Pliocene boundary. Assuming the F3 folds are structurally linked to the post-Miocene activity of the Taknar Fault, the mean θ angle (measured anticlockwise from the F3 fold trend) is 12°, compatible with a pure shear-dominated regime of sinistral transpression (Figures 13a-c). The N-S-directed maximum compression direction then favoured reactivation of the pre-existing Sebeh-Kait-e-Bargh and Taknar faults as sinistral transpressive faults (Figure 13c). This scenario is compatible with the seismotectonic scenario in the DF region (Farbod et al., 2011; Mousavi et al., 2013; Shabanian et al., 2010; Zamani et al., 2008) (Figure 2).

The final deformation event is recorded by the F4, NW-SE trending, folding episode, which locally affect and refoimd the F3 folds in the URF (Figures 3 and 9a). The F4 folds have been interpreted as the evidence of the transition from dextral to sinistral shearing along the DF, in consequence of a shift of the regional shortening direction from NW-SE to NE-SE-directed (Javadi et al., 2013). This interpretation is at odds with the structural evidence reported in this study and neighbour regions, where N-S shortening has been demonstrated to control the post-Miocene tectonic scenario at regional scale (Shabanian et al., 2010; Tadayon et al., 2017). In this view, an alternative hypothesis considers the genesis of the F3 folding as kinematically linked to the sinistral reactivation of the major NE-SW fault systems responsible for the NW translation of the tectonic block to the north of the Taknar Fault, segmented by the Dahan–Gaeh Fault and delimited by the Khartaran Fault (Figure 2).

7.1 | Tectono-stratigraphic scenario

Combining the structural, stratigraphic and tecthronomorphological dataset presented in this study with those available from the DF region (Calzolari, Della Seta, et al., 2016; Calzolari, Rossetti, et al., 2016; Calzolari et al., 2018; Javadi et al., 2013; Tadayon et al., 2017), we propose a tectono-stratigraphic evolutionary model for the DF (Figure 14). The model starts in the Upper Cretaceous, when NE-SW-oriented rift systems determined the breakup of the continental lithosphere during formation of the Sabzevar back-arc oceanic domain. At this stage, the Taknar and Sebeh faults were generated as basin boundary faults (Figure 14a). These acted as major weakness zones and their polyphase tectonic reactivation controlled the post-breakup tectono-stratigraphic evolution of the study area.

An early phase of positive basin inversion and a source-to-sink scenario is documented during the deposition of the early Palaeocene deposits of the Kerman Fm, formed at the expenses of the exhuming ophiolite units. This early phase of basin inversion occurred under a NW-SE-directed maximum shortening direction (Figure 14b). This phase is followed by a period of tectonic quiescence and regional subsidence during deposition of a thick pile of Palaeocene-Eocene volcaniclastic deposits (60–40 Ma; Figure 14c). Such volcanic successions are widely recognized across entire Iran with reported thicknesses of several kilometres (Berberian & King, 1981). These deposits caused the resetting of the AFT and AHe thermochronological
systems in most samples. To the east, this period corresponds to the genesis of KAIC that intrudes these volcanic-elastic succession (Shafai Moghadam et al., 2015), providing a further support to the relevant thickness of the host units.

Tectonic activity resumed in the Late Eocene, in consequence of renewed NW-SE shortening which caused re-activation of the Taknar and Sebeh faults and of the inherited Mesozoic basin-boundary faults to the east (Tadayon et al., 2017), together with folding of the volcanoclastic succession (Figure 14d). This renewed tectonic re-activation caused formation of a major restraining stepover (Cunningham & Mann, 2007) in consequence of the propagation of E-W dextral shearing across pre-existing oblique inherited crustal anisotropies (the NE-SW-oriented Taknar and Sebeh-Kalat-e-Bargh faults). This caused exhumation of the structural high bounded by the Taknar and Sebeh-Kalat-e-Bargh faults that was accompanied by erosion of the Late Palaeocene-Eocene volcanoclastic deposits. Such deposits constituted the source area for the Oligocene-Miocene sediments, deposited in the adjacent, structurally low domains. This time frame also corresponds to the syn-tectonic deposition of the LRF and Qom Fm in the northwestern sector of the study area, as attested by syn-sedimentary folding of these units (Figure 14e,f). During early Miocene times, an early phase of fault-related exhumation is documented to the south of the active trace of the DF (Cabolari, Rossetti, et al., 2016). Continuous shortening caused the nearly steady-state

TADAYON ET AL.

FIGURE 14 Conceptual, spatio-temporal tectono-stratigraphic model for the evolution of the DF region since the Upper Cretaceous (see text for further details) [Colour figure can be viewed at wileyonlinelibrary.com]
exhumation of the push-up horst bounded by the Taknar and Sebeh fault systems, providing the source area for the deposition of the URF in the adjacent regions. The thickness of the URF was laterally discontinuous. To the south of the DF, thicknesses were sufficient to reset the A-He system (Calzolari, Rossetti, et al. 2016) (Figure 14g).

At the Miocene–Pliocene transition, the regional stress field changed, resulting in a shift of the maximum compression direction (σ2) from NW–SE to NS (see also Tadayon et al., 2017). This new tectonic regime was responsible for the superimposed E-W-oriented folding, the generation of a conjugate sets of NW–SE dextral and NE–SW sinistral strike-slip faults, and the reactivation of the Taknar and Sebeh faults as sinistral faults. Major N–S extensional faults accommodate the E–W trending maximum extension (σ2) in the region, causing regional unroofing and exhumation. This tectonic regime is assumed to be presently active and to control the seismotectonic scenario of the region. In this view, the active DF is interpreted as a thrust-dominated, segmental, sinistral transpressive fault system. To the south, the Miocene–Pliocene tectonic reorganization is recorded by activation of the dextral NE–SW (Kuh Sahranj Fault; Nozaem et al. 2013) and E–W (Kuh Faghan Fault; Calzolari, Rossetti, et al. 2016; Calzolari et al. 2018) fault strands, whose activity is documented throughout the Quaternary times (Calzolari, Della Seta, et al. 2016) (Figure 14h).

7.2 | Tectonic implications

Our analysis documents a polyphase tectonic history of the DF region since the Upper Cretaceous. Notably, our reconstruction overlaps in time with the main tectono-stratigraphic events documented both along the Arabia–Eurasia convergent boundary and in its hinterland domain, providing the possibility to link the intraplate deformation in Central Iran with the spatio-temporal tectonic evolution at regional scale.

To achieve this purpose, we have collected the available AFT and AHe data from the main tectonic domains of the Iran region (Figure 1). A total of 197 AFT and 215 AHe ages were compiled from the literature (Axen et al. 2001; Ballato et al. 2013; Calzolari, Rossetti, et al. 2016; Doliar 2010; Francois et al. 2014; Gavilott et al. 2010; Guest, Axen, Lam, & Hassanzadeh, 2006; Guest, Stocki, et al. 2006; Homke et al. 2010; Kargaranbafghi et al. 2012; Khadivi et al. 2012;...

The major period of tectonic quiescence and regional subsidence is instead recorded during the early-middle Eocene that is consistent with the timing of regional extension, magmatism, and core-complex formation and exhumation in Central Iran (Figure 15).

The reconstructed scenario suggests a scenario of cyclic changes in plate coupling along the Zagros convergence margin through time that can be tentatively correlated with the interplay of advancing (Upper Cretaceous–Palaeocene and Late Eocene–Miocene and Miocene–Pliocene) and retreating (early-middle Eocene and middle Miocene) stages of the Zagros plate margin. Stresses transmitted from the plate boundary were responsible for the tectonic re-activation of structural anisotropies inherited from the post-Cimmerian fragmentation of the Gondwana supercontinent that guided and accommodated punctuated intraplate deformation in Central Iran since the Cretaceous times and throughout the Cenozoic.

The punctuated re-activation of the northwestern boundary of the CEIM as a major dextral (dominantly transpressional) shear belt throughout the Cenozoic (Bagheri et al., 2016; Calzolari, Rossetti, et al., 2016; Calzolari et al., 2016; Javadi et al., 2013; Nozaeem et al., 2013; Tadayon et al., 2017; this study) has significant implications on the tectonic reconstruction of the hinterland domain of the Zagros convergence zone. In particular, it imposes a profound re-interpretation of the proposed tectonic and geodynamic models of the region and requires a new class of geodynamic models able to incorporate the effects of the advancing and retreating stages of the Zagros plate margin in the residual stress distribution far away from the Arabia-Eurasia convergence zone.

**8 | CONCLUSIONS**

A new tectono-stratigraphic and T-t reconstruction for the DF region since the Cretaceous times is presented in this study. This allows to link the tectonic evolution in Central Iran with the spatio-temporal tectonic evolution at regional scale. The main results can be summarised as follows:

1. A long-lasting history of fault-related exhumation, burial and cooling starting in the Upper Cretaceous is documented for the DF region.
2. The DF region operated as a zone of residual stress accommodation and transfer in the hinterland domain of the Zagros convergence zone throughout the Cenozoic times. Its tectonic evolution was modulated by the plate-boundary dynamics and the modes of tectonic reactivation of the inherited intracontinental weak zones in Central Iran and at its tectonic boundaries.
3. The northern boundary of the CEIM was affected by a major and polyphase dextral deformation during the Cenozoic.
4. The Zagros convergence zone and its hinterland domain were fully mechanically coupled since ca. 40–35 Ma, a time lapse that is here referred to as the onset of continental collision along the Arabia-Eurasia plate margin.
5. At the Miocene–Pliocene boundary, major change in tectonic regimes is recorded within the DF region, as a consequence of a...
shift in the regional maximum compression direction from around NW-SE to NS. This corresponds to a major regional re-organisation of the plate boundary, following the transition from an infant to a mature stage of continental collision in the region.

6. The present trace of the DF represents the last shear deformation increment of a polyphase and long-lasting zone of intraplate deformation situated at the northern edge of the Lut Block of Central Iran.

Finally, the interplay of advancing and retreating of the convergent plate margin may be proposed as the main cause of the intraplate deformation in the hinterland domain of the Arabia-Eurasia convergence zone during Mesozoic-Cenozoic times. These results further support the potential of the intraplate deformation zones to be used as a proxy to reconstruct the long-term dynamics and evolution of the deforming plate margins.

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SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section at the end of the article.


APPENDIX A

Apatite fission track thermochronology (AFT)

The apatite grains were recovered from the collected samples following standard crushing, sieving, washing, magnetic and heavy liquid separation. Apatite grains were mounted in Aralite® adhesive, ground, and polished to expose internal mineral surfaces. Etching with 5.5 M of HNO₃ at room temperature for 20 s revealed spontaneous fission tracks intersecting the apatite surface. Samples were covered with a uranium-free muscovite external detector and irradiated with thermal neutrons at the Radiation Center of the Oregon State University. Induced fission tracks in the external detector were revealed by etching the mounts in 40% HF at room temperature for 40 min. The fission tracks were counted by the first author with Zeta number of 362.48 ± 12.69, under a nominal magnification of 1250× on an Olympus BX51 equipped with an automatic stage and a video camera (at the Thermochronology Laboratory, Department of Geosciences, University of Padova). The single grain age distribution of each sample was decomposed into age populations that fit the measured distribution by using the best-fit binomial peak fitting method (Galbraith & Green, 1990). The Trackpy 4.2 program was used for all AFT age calculations procedures (Dunkl, 2002). A chi-squared (χ²) test is carried out on the AFT single grain age in order to test the homogeneity of data (Galbraith, 1981). The probability of χ² is calculated for each sample; if P(χ²) > 5% then the sample is assumed to be homogenous (Galbraith & Laslett, 1993).

Apatite (Uranium–Thorium)/Helium thermochronology (AHe)

The AHe analyses were carried out at Arizona University. Euahedral apatites were picked using a cross-polarized binocular microscope. Most grains had a minimum diameter of 90 μm and were inclusion free to avoid effects of He-implanting from inclusions or excess loss of He during decay due to a large surface/volume ratio (Farley, 2000). The grain dimensions were measured for calculation of the alpha-ejection (Fe) correction factor after Farley, Wolf, and Silver (1996), and single grains were packed in Nb-tubes for U-Th/He measurement. For each sample, three aliquots were prepared for analysis in order to ensure sample age reproducibility. The concentration of ⁴He was determined by the ³He isotope dilution and measurement of the ⁴He/³He ratio through a quadrupole mass spectrometer. Apatite samples were heated for 5 min at 11 A with a 960-nm diode laser for degassing. Each sample was reheated and measured to ensure that all gas was extracted in the first run. U, Th concentrations were obtained by isotope dilution using an inductively coupled plasma mass spectrometer. The HeFTy 1.9 program was used for AHe thermal modelling (Ketcham, 2005).