Cenozoic temporal variation of crustal thickness in the Urumieh-Dokhtar and Alborz magmatic belts, Iran

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Abstract

We present regional variations of whole-rock Sr/Y and (La/Yb)N ratios of magmatic rocks along the Cenozoic Urumieh-Dokhtar and Alborz magmatic belts, Iran. Both the magmatic belts are located at the north of the main Zagros Neo-Tethyan suture. The Urumieh-Dokhtar magmatic belt (UDMB), which trends NW-SE for 1000 km across Iran, was characterized by the intensive volcanism and plutonism, and defined the magmatic front (MF) of the Zagros orogenic belt. The Alborz magmatic belt (AMB) is situated to the north, and characterized by less intense magmatic activity. The Alborz magmatic belt was formed behind it in the rear-arc (RA) domain. A striking feature of the both magmatic belts is the transition from normal calc-alkaline arc magmatism during the Eocene–Oligocene to adakite-like calc-alkaline magmatism during the Middle to Late Miocene–Pliocene. The late-Cenozoic magmatism of the UDMB and AMB shows higher Sr/Y and (La/Yb)N. However, it should be noted that crustal thickening event is intensive in the UDMB than AMB during Late Cenozoic. Using the composition of the Lale-Zar zircons from the SE UDMB we determined the oxygen fugacity (fO2) during zircon crystallization to be between FMQ (fayalite–magnetite–quartz buffer) -0.69 to +2.41, whereas those of the Hashroud-Teckmdash-Gormolla zircons from NW AMB range from −1.22 to +5.99. The fO2 estimates suggest relatively more oxidized conditions for the Late Cenozoic igneous rocks of the AMB. Compiled data from the UDMB and AMB intrusions show an increase in average zircon crystallization temperatures with decreasing age. These outcomes have been interpreted in terms of variation of the crustal thickness, from 30 to 35 km during Eocene-Oligocene to 40–55 km during the middle-late Miocene. We propose the increase in crustal thickness is associated with the collision between the Arabian plate and Iran and subsequent convergence during the middle-late Miocene.

Key words: Crustal thickness, Oxygen fugacity, Alborz magmatic belt, Urumieh-Dokhtar magmatic belt

1. Introduction

Subduction zone magmatism and other plate-margin processes that occur within accretionary orogenic belts throughout plate convergence provide a record of continental crust formation and/or crustal thickening (i.e., Robinson et al., 2014). The crustal thickness and depth of melting of magmatic rocks can be estimated based on major and trace element contents (i.e., SiO2, rare earth elements (REE), Sr) or their ratios t (i.e., K2O/Na2O, (La/Yb)N, Ce/Y, Sr/Y) (Profeta et al., 2015). For example, Sr is favorably incorporated into plagioclase at low pressures (≤1.0 GPa) during partial melting and magmatic fractionation (Rapp et al., 2003). At higher pressures (≥1.2 GPa), plagioclase is destabilized in favor of garnet, leading to a Sr enrichment of the melt phase. Also, garnet is supposed to be stable at high-pressure, so melts will be depleted in garnet-compatible elements, such as Y, Sc, and heavy REE (HREE) (e.g., Hernández-Uribe et al., 2020). Consequently, elevated Sr/Y and (La/Yb)N ratios of igneous rocks records high-pressure partial melting and fractionation, and thus an increased crustal thickness (Chapman et al., 2015; Profeta et al., 2015). Such correlated geochemical and crustal thickness variations were reported in the western North American Cordilleran interior (Chapman et al., 2015), Himalaya-Tibet (Laske et al., 2013), Urumieh-Dokhtar magmatic belt (Iran) (Chaharlang et al., 2020) and the eastern Sakarya Zone, NE Turkey (Karsli et al., 2020). Furthermore, oxygen fugacity estimations may also provide suitable proxies for the crustal thickness (e.g., DePaolo et al., 2019) in the orogenic belts where the crust has a distinct oxygen fugacity from the underlying mantle.

This paper focuses on the temporal variation of the crustal thickness in the Urumieh Dokhtar and Alborz magmatic belts (Iran) throughout the Cenozoic, on the basis of whole-rock and zircon geochemistry. Our results indicate that the Eocene to Oligocene igneous rocks of the UDMB are normal calc-alkaline arc magmas, and those of the AMB and Middle to Late Miocene–Pliocene igneous rocks of the UDMB are calc-alkaline/shoshonitic and adakite-like calc-alkaline igneous rocks, respectively. This change probably reflects an increase in crustal thickness during the Middle to Late Miocene–Pliocene. We discussed below whether this increase in crustal thickness are related to ongoing convergence between the Arabian plate and Eurasia after the initial continental collision or not.

2. Geological Framework

The Cenozoic magmatic belts in Iran can be separated into the four sub-belts: (i) the Urumieh–Dokhtar magmatic belt (UDMB) (~60–5 Ma) (Fig. 1a and b), (ii) the Alborz magmatic belt (AMB) (~60–5 Ma) (Fig. 1a and c), (iii) the east Iranian magmatic belt (EIMB) or Gonabad–Birjand–Zahedan magmatic belt (~50–5 Ma) and (iv) the Makran magmatic belt (40–5 Ma) of southeast Iran (Fig. 1). Their development has been related to subduction and collision along the Neo-Tethyan suture zones and subsequent convergence between Eurasia and Arabian plates (e.g., Berberian and King, 1981). The Cenozoic intermediate to felsic magmatic rocks dominate along the UDMB, with few mafic rocks (Fig. 1d), whereas Late Cenozoic (15–5 Ma) mafic rocks are more abundant in the AMB (Fig. 1e).

The igneous rocks with shoshonitic affinity are mainly limited to the AMB regions (Fig. 1e; Castro et al., 2013; Moghadam et al., 2018), where the magmatism was partly associated with the extension and thinning of continental crust (Sepidbar et al., 2019). In this study, we focus on magmatism along the Urumieh–Dokhtar magmatic and the Alborz magmatic belts to decipher the systematic variations crustal thickening and sources of magmas in both belts over the time period from 80 to 5 Ma. The general stratigraphic and magmatic characteristics of the both belts are summarized below.

2.1. The Urumieh–Dokhtar magmatic belt (UDMB)

The Urumieh-Dokhtar Magmatic Belt is 50–80 km wide, defines the magmatic front of the Cenozoic Iran arc and trends NW-SE for 1000 km across Iran between 28° and 39°N (Fig. 1a) (Fig. 1; Sepidbar et al., 2019). The magmatic belt is built over a basement comprising, from the base to top, (i) The Eocene–Oligocene magmatic rocks with a geochemical signature typical of a continental arc under extension regime from ~55 Ma until ~37 Ma (e.g., Verdel et al., 2011), (ii) The Oligocene oceanic island basalt (OIB)-like magmatism, indicating a transition from arc-like to “OIB-like” (Sepidbar et al., 2019); (iii) the Late Oligocene–Miocene arc magmatism induced by the beginning of collision (e.g., Topuz et al., 2019); and (iv) the Late Miocene–Quaternary magmatism, which formed in response to the onset of closure of the Tethys ocean. However, it should be mentioned that the Middle Eocene magmatism volumetrically dominates over all others (Verdel et al., 2011).

A late-Paleocene to middle-Eocene marine transgression is represented by thick (500–1000 m) sequences of the deep marine Nummulitic limestones (Fig. 2a). These limestones show variation in thickness and often grade into fine-grained pyroclastic rocks. The Eocene flare-up magmatism in the UDMB coincides with the Eocene transgressions, which occurred during a rapid fall in global sea level (Miller et al., 2005). In this period, the magmatism was associated with extension-related subsidence (Fig. 2a), which allowed for the presence of the submarine magmatism, as has been described by Hassanzadeh (1993). This unit is typically overlain by the late-Eocene–early-Oligocene Lower Red Formation that is composed of conglomerate, sandstone, shale, and gypsum, as well as relatively limited pyroclastic deposits and volcanic flows. A late-Eocene to late-Paleocene regression resulted in deposition of red volcano-sedimentary sequences (Fig. 2a).

The Oligocene to Early Miocene marine transgression series of limestones and marls interlayered with mafic lava flows that are known as the Qom Formation (~1200 m), cover the Lower Red Formation (Reuter et al., 2009). The Qom Formation is conformably overlain by the Miocene Upper Red Formation and Pliocene to Quaternary continental sediments (Fig. 2a).

Generally, the UDMB is subdivided into central and southeast segments. The central segment is located between the Saveh and Yazd cities, and includes the Qom-Kashan-Natanz, Ardestan, Nain, and Nain–Yazd domains (Fig. 1). This segment comprises the Late Neoproterozoic to Paleozoic sedimentary rocks overlain by the Mesozoic series, which are in turn covered by Paleogene volcanic rocks. The whole sequence is locally intruded by the Oligocene to Miocene gabbroic and granitic plutons (Berberian and King, 1981). The southeastern segment is characterized by three different sequences (e.g., Hassanzadeh, 1993). The early Eocene Bahraseman complex comprises 5.5-km-thick acidic pyroclastics, tuffs, volcanic breccias and trachybasaltic to trachyandesitic lava flows. The Middle to Late Eocene Razak complex is composed of ~7.5 km-thick basaltic–rhyolitic volcaniclastic sequences (ca. 37.5 ± 1.4 Ma; Hassanzadeh, 1993). The Oligocene Hezar complex is made of 1.3 km thick trachyandesite and trachybasalt. The Arsingle bondAr analcime dating yielded an age of 32.7 ± 6.3 Ma (Early Oligocene) for this unit (Hassanzadeh, 1993). The Eocene to Oligocene volcanoclastic and associated intrusive rocks are unconformably overlain by the Late Oligocene to Miocene continental red beds and limestones. Such formations reveal that that the area had emerged above sea level.

The youngest subvolcanic and volcanic rocks along the southeast UDMB are composed of Late Miocene–Pliocene dacites, basalts and foidites which occur in the form of domes and lava plugs. The Masahim stratovolcano has an Arsingle bondAr biotite age of 6.8 ± 0.4 Ma (Hassanzadeh, 1993) and is the prominent example of recent magmatic activity. In this contribution, we studied units from the Lale-Zar region, the southeast UDMB, and compiled our results with other parts of the southeast UDMB.

2.2. The Alborz magmatic belt (AMB)

The Alborz magmatic belt is ~600 km long, ~100 km wide, and extends in a roughly E–W direction (Sepidbar et al., 2019; Fig. 1). During the Mesozoic, extension in the central Alborz led to upper Late Triassic rift volcanism (Rhaetic) and deposition of shaly and coal-bearing Shemshak Formation. These units are overlain by the Upper Jurassic and Lower Cretaceous shallow marine limestones interlayerd with alkaline basalts (Barrier et al., 2018). This is known as the Chalus Formation. The Mesozoic volcano-sedimentary rocks are unconformably overlain by the Paleocene to Eocene volcanic rocks and their pyroclastic deposits. The Eocene to Oligocene magmatism is represented by shoshonitic and calc-alkaline mafic to felsic magmas, while the Miocene-Quaternary magmatism is mostly high-K alkaline and adakitic in composition, with minor sodic alkaline bodies (e.g., Aghazadeh et al., 2011; Verdel et al., 2011). This area was subaerial, as deduced from the deposition of red volcano-sedimentary sequences from the Middle- to the Late-Paleocene (Verdel et al., 2011). A major marine transgression during the Early to Middle Eocene is documented by up to 1000 m thick pile of marine limestones and magmatism induced by extension (Fig. 2b). The Eocene–Oligocene boundary is related to the final closure of the Neo Tethys oceanic corridor in the Caucasus (e.g., Barrier et al., 2018) and the Eastern Sakarya Zone (Karsli et al., 2020). This major compressional stage marked the transition from back-arc extension to collisional tectonics in this region (Francois et al., 2013; van der Boon et al., 2018). The Early–Middle Miocene corresponds to a period of transition from marine to continental sedimentation (Fig. 2b) in the Iranian plateau and a major change in the magmatic activity in the region, from dominantly calc-alkaline to high-K calc-alkaline in composition (Sepidbar et al., 2019). A further major episode of intracontinental shortening and regional exhumation also occurred during the Early–Middle Miocene, as documented along the Bitlis–Zagros collisional zone. In this contribution, we studied the Hashroud, Gormolla and Techmedash intrusions situated in the northwest of the AMB, and compared results with a compilation of data obtained from other parts of the northwest AMB.

3. Field relations and petrography

We collected representative samples from plutonic rocks from Lale Zar and Hashroud-Goemolla-Techmedash from SE UDMB and NW AMB, respectively, for geochronological and geochemical analyses. Below we discuss these rocks in more detail.

3.1. Oligocene granitoids from the southeast UDMB

We considered the Oligocene granitoids of Lale-Zar, dated in this study, from the southeast UDMB. The Lale-Zar granitoid intruded into the country rocks composed of the Late Cretaceous limestone with local sandstone interlayers and the Eocene volcanic and volcano-sedimentary rocks, and the pluton crops out over an area of ~120 km2. Despite its large size, the intrusion is relatively homogenous and is petrographically made of monzonite, monzodiorite, and diorite-granodiorite.

All the intrusions occur as elongated bodies (less than 5 km) within one stock bounded by the Eocene volcanic and pyroclastic rocks in the south, whereas it abuts Miocene–Pliocene sedimentary rocks in the north. The contacts between the intrusion and the Eocene volcanic sequences are predominantly sharp and discordant (Fig. 3a). Generally, these intrusive rocks are not deformed and do not show alteration nor weathering. Granodiorite contains mafic microgranular enclaves (MMEs), commonly 2–5 cm in size. The enclaves show curved shapes. Granodiorites display a granular to microgranular porphyritic texture, and contain plagioclase (40–45%), K-feldspar (15–25%), quartz (20–24%), biotite (5%), and hornblende (5%). Zircon, magnetite, ilmenite, apatite, and titanite are accessory minerals (Fig. 3b). Monzonite/monzodiorite are subordinate, and occurs in the center of the stock whereas granodiorite and diorite are exposed near the northern and southern margins. They differ from granodiorite by modal compositions of K-feldspar (20%–42%), quartz (1%–5%), plagioclase (~10%–35%), and hornblende and/or biotite (4%–5%) (Fig. 3c). Diorites are grayish and porphyritic and are characterized by abundant subhedral to anhedral plagioclase (~65%), amphibole and biotite (10%–20%), quartz (<5%) and rare pyroxene. Zircon, apatite, titanite and opaque minerals are accessories. Plagioclase contain small inclusions of euhedral to subhedral biotite. It is partly altered to sericite and carbonate. Rare clinopyroxene occurs as fine-grained interstitial crystals between feldspars and as inclusions in feldspar. Subhedral to anhedral amphiboles occur as fine- to medium-grained phenocrysts. Anhedral quartz is interstitial between feldspars.

3.2. Eocene to Miocene intrusions of the northwest AMB

We considered the Eocene to Miocene Hashroud, Gormolla and Techmedash intrusions, dated by Sepidbar et al. (unpublished data), from the northwest AMB. The Hashroud intrusion is exposed over an area of ~35 km2, intruding into the Eocene volcanoclastic rocks, conglomerate, volcanic ash and pumice strata (Sepidbar et al., unpublished) (Fig. 3d). The intrusion is made of medium- to coarse-grained granite. Granites are characterized by porphyritic textures, and contain quartz (30–35%), plagioclase (32–35%), K-feldspar (28–30%), and biotite (5%) (Fig. 3e). Magnetite, zircon, apatite, and titanite are accessory minerals. Subhedral to anhedral plagioclase (0.5–2 mm) is altered to clay minerals and sericite, and show compositions ranging between oligoclase and albite. Quartz grains are anhedral and/or interstitial between plagioclase and K-feldspar.

The Techmedash intrusion is located at the northwest segment of the Alborz magmatic belt, and is exposed over an area of ~30 km2. It was intruded into Eocene volcanoclastic rocks, including andesite, trachyandesite, alkali basalt and pyroclastic rocks. The Techmedash intrusion is mainly composed of gabbro-diorite, which contains plagioclase (~60%) and clinopyroxene (~30%), ± olivine ± biotite. Apatite and titanomagnetite are common accessories. Sericite, epidote, chlorite, calcite and titanite are secondary phases (Fig. 3f).

The Gormolla intrusion is made of tonalite, intruding into the Eocene andesite, trachyandesite, alkali basalt and pyroclastic rocks. Tonalite includes plagioclase (40%–45%), quartz (20%–25%), k-feldspar (10%–15%) and hornblende and biotite (5%–10%). Plagioclase is 0.2 to 3.0 mm in diameter and is subhedral to anhedral. Alkali feldspar laths vary from 0.5 to 2.5 mm in diameter. Biotite and hornblende show slight alteration to chlorite. Gabbro is composed of plagioclase and clinopyroxene along with minor quartz. Plagioclase is altered into sericite and titanite and pyroxenes are chloritized. Zircon, apatite and iron oxide minerals occur as accessories.

Analytical methods

The compiled data come from several intrusive and volcanic rock of Eocene to Miocene ages in the UDMB (Supplementary Table S1a) and AMB (Supplementary Table S1b) for whole rock major and trace-element geochemical composition.

Fifty samples were collected from the Lale-Zar (Supplementary Table S2a) and Hashroud-Gormolla and Techmedash intrusion (Supplementary Table S3). The least-altered samples were then selected for bulk-rock geochemical analysis. Whole-rock major and trace element concentrations were analyzed by means of an ARL Advant-XP automated X-Ray Fluorescence (XRF) spectrometer hosted at the Department of Physics and Earth Sciences of the University of Ferrara, Italy. Accuracy and precision were better than 2–5% for major elements and 5–10% for trace elements. Detection limits were 0.01 wt% for major elements and 1–3 ppm for trace elements concentrations, respectively. Rubidium, Sr, Y, Nb, Hf, Ta, Th, U, and REE (analyses) were carried out by means of a Thermo Series X inductively coupled plasma-mass spectrometer (ICP-MS) hosted at the Department of Physics and Earth Sciences of the University of Ferrara, Italy. Precision and accuracy were better than 10% for all elements, well above the detection limits (see Casetta et al., 2020 and references therein).

For Usingle bondPb zircon dating, zircons were separated from two monzonites (Laz-4, and Laz-11) and one monzodiorite (Laz-7) and one diorite (Laz-26) (Supplementary Table S2b). Zircon separation has been performed by conventional techniques including crushing by hammer, sieving, magnetic and heavy liquid separations. Separated grains and the 91,500 zircon standard were mounted in epoxy. The obtained mounts were then polished to expose the zircon cores for analysis. All the zircons were examined via transmitted and reflected light micrographs and cathodoluminescence (CL) images to reveal their internal structures =. Isotopic data were acquired by laser ablation inductively coupled plasma-mass spectrometry (LA-ICP-MS) at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGGCAS) in Beijing, China. Analyses were performed using an Agilent 7500a quadrupole ICP-MS and a Thermo-Finnigan Neptune multi-collector ICP-MS connected to a 193 nm Excimer ArF laser-ablation system (Geolas plus). Operating conditions included a beam diameter of 32 μm, 6 Hz repetition rate, and ablation pit depth of ~40 μm. Helium was used as a carrier gas to enhance transport efficiency of the ablated material. Reference zircon 91,500 was used for calibration of age determination in all cases, that is used for all zircons by different ages, and reference zircon NIST SRM 610 was used as the external standard for trace element concentration determination. Trace element concentrations were calculated using GLITTER 4.0 and calibrated using 29Si as an internal standard. However, it should be mentioned that no standard treated as unknowns. The detection limit of 238U for SRM 610 is 14,000 cps/ppm using a spot size of 40–60 μm and 60 μm for the standards with laser repetition of 10 Hz and laser energy density of 25 J/cm2. The mass stability during analysis was better than 0.05 amu/24 h. The weighted mean Usingle bondPb ages and inverse-concordia plots were produced by Isoplot/Ex v. 3.0 (Ludwig, 2003). Usingle bondPb inverse-concordia (207Pb/206Pb vs 238U/206Pb) ages and trace and rare earth elements for zircons from the Lale-Zar magmatic rocks are given in Supplementary Table 2b and Supplementary Table 2c, respectively.

Of the many models that can calculate the zircon saturation temperature (TZircsat.), TZircsat. of magmatic rocks from the UDMB and AMB were calculated based on calibration of Watson and Harrison (1983) (Eq. (1)), which was revised by Boehnke et al. (2013) (Eq. (2)) (Supplementary Table S4a and Supplementary Table S4b). These equations have been widely used to to estimate the peak temperature experienced by magmatic rocks (i.e., accessory mineral thermometry; Watson and Harrison, 1983).

Moreover, crystallization temperatures of magmatic rocks from the Lale-Zar and Hashroud-Techmedash-Gormolla regions of the UDMB and the AMB, respectively, were estimated using the Ti-in-zircon thermometer (Eq. (3)) from the equations of Ferry and Watson (2007) (Supplementary Table S4c), whose calibration assumes crystallization under rutile- and quartz-saturated conditions (aTiO2 = 1; aSiO2 = 1) at 10 kbar.

For the investigated samples, T is Kelvin (Ferry and Watson, 2007), aSiO2 was assumed to be unity, as all systems reached quartz saturation, and aTiO2 was calculated using the TiO2 solubility model of Hayden and Watson (2007). Though present, rutile has not been reported as part of the magmatic mineral assemblage and is generally considered to have a hydrothermal origin. Thus, aTiO2 = 0.6 has been used in the calculations in agreement with the general presence of other Fesingle bondTi oxides (e.g. Ferry and Watson, 2007).

Oxygen fugacity of magmatic rocks from the Lale-Zar and Hashroud-Techmedash-Gormolla regions of the UDMB and the AMB, respectively, were estimated using values of Dzircon/meltCe and the REE contents in zircon (Eqs. (4), (5), (6); Supplementary Table 4c).

Partition coefficients for the trivalent REEs and Hf^{4+}, Th^{4+}, and U^{4+} are used to calculate $D_{zircon/melt}Ce^{3+}$ and $D_{zircon/melt}Ce^{4+}$, respectively (Blundy and Wood, 1994). These authors proposed that the mineral-melt partition coefficient for cation i can be related to the lattice strain energy created by substituting a cation whose ionic radius (r_i) differs from the optimal value for that site (r_0).

A plot of ln D_i versus $(r_i/3 + r_0/6)(r_i - r_0)^2$ produces a linear trend for isovalent cations (Smythe and Brenan, 2016). Partition coefficients of Ce^{3+} and Ce^{4+} can thus be evaluated by interpolation. As Ce is a combination of Ce^{3+} and Ce^{4+}, the value of $D_{zircon/melt}Ce$ range between these two partition coefficient end-members, and the following equation (Eq. (6)) is used to evaluate the fO_2 during magma fractionation, where T can be calculated by Ti-in-zircon thermometer and the unit is Kelvin (Ferry and Watson, 2007), NBO/T is the proportion of non-bridging oxygens to tetrahedrally coordinated cations (Virgo et al., 1980) calculated on an anhydrous basis (assuming all ferrous iron), and xH_2O is the mole fraction of water dissolved in the melt and T is in Kelvin and can be determined using the Ti-in-zircon thermometer (Ferry and Watson, 2007). Zircon/melt partition coefficients for Ce^{3+} ($D_{zircon/melt}Ce^{3+}$) and Ce^{4+}($D_{zircon/melt}Ce^{4+}$) can be calculated for individual zircon-melt pairs using the lattice strain model of Blundy and Wood (1994), and are constrained by the partition coefficients for the other REEs as well as Hf, Th, and U. Here, we omitted the LREEs, La and Pr, in the appropriate formula for trivalent cations, since they are very minor components in zircons, and absorptions are extremely vulnerable to contamination by accessory mineral inclusions. Eu exists as both $Eu^{2+}$and Eu^{3+}species, although the former is sensitive to plagioclase crystallization, therefore Eu was excluded. Values of r_0 were 0.93 Å (determined by regression) and 0.83 Å (8-fold coordinated Hf) for 3+ and 4+ cations, respectively. Propagation of error through the fO_2 calculation, including the error in Ce^{3+}/Ce^{4+} in the melt [Eq. (6))], Ti crystallization temperatures (Eq. (3)), analytical errors in the calculated NBO/T as well as xH_2O, yield an estimated uncertainty in the accuracy of ~1 log unit (Supplementary Table 4c).

The Sr/Y (Eq. (7)) and $(La/Yb)_N$ (Eq. (8)) ratios are often used to estimate the depths of magma crystallization (Chaharlang et al., 2020; Chiaradia, 2015). The experimental equations relating trace element ratios to crustal thickness (Profeta et al., 2015) used here

5. Results

Whole-rock geochemistry

Late Cretaceous–Late Eocene magmatic rocks (60–30 Ma) from the UDMB comprise foid bearing gabbro, foid bearing monzo-gabbro, foid monzo-syenite, gabbro, gabbroic diorite, diorite, tonalite/granodiorite and granite (Fig. 4a), while those from the AMB include gabbro, gabbroic-diorite, monzodiorite, monzonite, quartz monzonite and syenite with less granite and tonalite/granodiorite (Fig. 4b). This suggests that the plutonic rocks from the UDMB are characterized by much more dispersed composition at equivalent silica contents compared to the AMB plutonic rocks. This is shown by A/CNK vs. ANK and SiO_2 vs K_2O diagrams

(Peccerillo and Taylor, 1975), where the AMB plutonic rocks have a more peraluminous and potassic signatures, relative to the UDMB igneous rocks (Fig. 4c-f).

The Early-Oligocene to Early-Miocene (30–15 Ma) UDMB magmatic rocks consist of gabbro, diorite, tonalite/granodiorite, with minor granite (Fig. 4a), while those from the AMB comprise monzodiorite, monzonite, quartz monzonite, syenite, foid-bearing gabbro, foid monzograbbro and foid monzosyenite and foid-bearing syenite (which are equivalents of basanite and phonolite–tephrite) (Fig. 4b). This shows that the Early-Oligocene to Early-Miocene magmatism from the AMB is characterized by more under-saturated compositions. These magmatic rocks from the AMB belong to high-K calc-alkaline and shoshonitic series, while those from the UDMB exhibit a low- and medium-K calc-alkaline affinity. All the magmatic rocks from the both belts show I-type character with the alumina saturation indices [ASI = Al/(Ca - 1.67P + Na + K)) (Frost et al., 2001) ranging from 1.1 to 1.9 for the UDMB and from 0.7 to 1.5 for the AMB, indicative of metaluminous to strongly peraluminous affinity.

The Early-Oligocene (29–23 Ma) intrusive rocks from the Lale-Zar in the southeast UDMB consist of gabbro-diorite, monzodiorite, monzonite and granodiorite (Fig. 4a; supplementary Table S2a), which is similar to other parts of the UDMB. These magmatic rocks belong to the low- and middle-K calc-alkaline series and have a metaluminous affinity (ASI = 0.7 to 1.1). A middle-Eocene intrusion (38.71 ± 0.95 Ma, Sepidbar et al., unpublished data) from the Hashroud in the northwest AMB (Supplementary Table S3) is granitic, with high K calc-alkaline to shoshonitic signatures (Fig. 4b and d). Early-Oligocene (~30 Ma) to Late-Miocene (~16 Ma) (Sepidbar et al., unpublished data) intrusions from the Gormolla and Techmedash regions, in the northwest AMB have granitic and gabbroic compositions, with high K calc-alkaline signature.

The REE chondrite-normalized patterns of Mesozoic to Middle Cenozoic magmatic rocks of the UDMB are characterized by a fractionation of the LREE relative to HREE ($(La/Yb)_N$ = 1.1 to 47) along with negative to slighlty positive Eu anomalies (Eu/Eu* = 0.4 to 1.1; av. 0.8), while those from the Late Cenozoic magmatic rocks of the UDMB are characterized by a similar fractionation of the light REEs (LREEs) relative to HREE ($(La/Yb)_N$ = 3.1 to 47), with smaller Eu anomalies (Eu/Eu* = 0.7 to 1.1); av. 1.0) (Fig. 5a-d). Chondrite-normalized REE patterns for the Lale-Zar intrusions in the southeast UDMB show similar trends with respect to other magmatic rocks from the UDMB, with enrichment in LREE compared to HREEs ($(La/Yb)_N$ = 3.1 to 13) and negative to slightly positive Eu anomalies (Eu/Eu* = 0.6 to 1.1; av. 0.86) (Fig. 5d).

The Early to Middle Cenozoic magmatic rocks of the AMB are characterized by fractionation of the LREE relative to HREE ($(La/Yb)_N$ = 3.0 to 58) along with negative Eu anomalies (Eu/ Eu* = 0.26 to 1.1; av. 0.8), while those from the Late Cenozoic magmatic rocks of the AMB are characterized by similar fractionation of the LREE relative to HREE ($(La/Yb)_N$ = 3.1 to 54), with s Eu anomalies (Eu/Eu* = 0.6 to 1.1; av. 0.9) (Fig. 5a-d). The Eocene to Oligocene magmatic rocks (~ 40 to 30 Ma; Sepidbar, unpublished data) of the Hashroud and Gormolla from AMB are characterized by fractionation of the LREE relative to HREE ($(La/Yb)_N$ = 11 to 33) along with negative Eu anomalies (Eu/Eu* = 0.7 to 0.9 (Supplementary Table S3)), which is similar from other Early-Eocene to Late Oligocene (60 to 30 Ma) intrusive rocks of AMB on the chondrite normalized patterns (Fig. 5). In addition the Miocene (16 Ma; Sepidbar, unpublished data) magmatic rocks of the Techmedash from AMB are also characterized by fractionation of the LREE relative to HREE ($(La/Yb)_N$ = 9.2 to 13.1), with Eu anomalies (Eu/Eu* = 0.7 to 0.8) similar to other Miocene of AMB (Fig. 5; Supplementary Table S3).

Zircon chemistry and U–Pb ages

Zircons from the UDMB intrusion are characterized by transparent to translucent prismatic grains showing oscillatory and sector zoning (Babazadeh et al., 2017; Haghighi Bardineh et al., 2018; Kananian et al., 2014; Kazemi et al., 2019; Nouri et al., 2020; Raeisi et al., 2019; Sarjoughian and Kananian, 2017). They range in length from ~100 μm to 300 μm, with length/width ratios of 1–3. Some grains have inherited cores (Fig. 6c). All analyses were screened by CL and BSE images to guide the choice of analytical spots and for the presence of possible inclusions which were measured together. The most of chondrite-normalized REE patterns are characterized by a steep increase from La to Lu with strong to moderate negative Eu anomaly (Eu/Eu* = 0.08–0.64), typical of magmatically crystallized zircons (Hoskin and Ireland, 2000). However, they are characterized by lower ∑REE than those of other units in the AMB (Fig. 7). A slight increase in ∑REE is observed with decreasing age, especially within the SE UDMB and NW AMB (Supplementary Table S1c). However, zircons from intrusive rocks of the southeast UDMB, NW AMB and NE AMB are more enriched in HREEs than those of the central UDMB (Supplementary Table S1c). Overall, zircon REE chondrite-normalized patterns and concentrations are typical of those reported for crustal zircons in general.

Zircon show U and Th contents that range from 34 to 935 ppm and from 15 to 1281 ppm, respectively for the Lale-Zar in the southeast and northwest UDMB. These zircon grains show Th/U ratios varying between 0.4 and 1.6, regardless of the investigated domain (Supplementary Table S2b). The Usingle bondPb zircon data indicate Oligocene weighted mean ages of 29.07 ± 0.84 Ma (Laz-26) to 24.8 ± 1.1 Ma (Laz-4), 23.64 ± 0.5 Ma (Laz-11), and 23.4 ± 1.3 Ma (Laz-7) on the inverse-concordia diagrams (Fig. 8; Supplementary Table S2c)..

Magma temperatures

Zr saturation temperatures

Whole-rock geochemistry was used to calculate TZircsat. of magmas (Siégel et al., 2017). Temperature based on zircon saturation (TZircsat.) and SiO2 contents are important to determine whether a magma that crystallized zircons was Zirc-undersaturated, zircon-saturated, or zircon-oversaturated.

Calculated TZircsat of late-Cretaceous to late-Eocene, Oligocene, and early-Miocene to late-Miocene magmatic rocks from the UDMB using Eq. (1) range from ~710 to 1013 °C mean ~ 864 °C), from 735 to 1041 °C (mean ~ 885 °C), and from 745 to 901 °C (mean ~ 890 °C), respectively (Table 1). The more recent model of Boehnke et al. (2013) yield wider TZircsat of ranges from ~721 to 1076 °C (mean ~ 896 °C), from 721 to 1134 °C (mean ~ 915 °C), and from 734 to 931 °C (mean ~ 877 °C), for the Late Cretaceous to Late-Eocene, Oligocene, and Early-Miocene to Late-Miocene magmatic rocks, respectively (Supplementary Table S4a). The model of Watson and Harrison (1983) (Eq. (1)) yields TZircsat of the Eocene, Oligocene and Early-Miocene to Late-Miocene magmatic rocks from the AMB range from ~852 to 1033 °C (mean ~ 945 °C), from 652 to 966 °C (mean ~ 840 °C), and from 862 to 1079 °C (mean ~ 946 °C), respectively. The more recent model of Boehnke et al. (2013) (Eq. (2)) yield ranges of ~870–1122 °C (mean ~ 1000 °C), 617–1037 °C (mean ~ 860 °C), and 925–1225 °C (av. ~1000 °C), respectively for same rocks (Table 1; Supplementary Table S4b).

Ti-in-zircon thermometry and oxygen fugacity

As mentioned above, zircons, that are selected from the UDMB and AMB intrusions, are characterized by transparent to translucent prismatic grains showing oscillatory and sector. Their magmatic character allows them to be used for crystallization temperatures (Eq. (3)) and oxygen fugacity (Eq. (4), 5, 6) of magmatic rocks.

The calculated Ti-in-zircon temperatures for the Early Oligocene magmatic rocks of the UDMB vary from 785 °C to 1141 °C (mean of 930 °C) for a total range in Ti concentration of 1.2–12 ppm (Table 1; Supplementary Table S4c; thermometer defined by Ferry and Watson (2007)). The calculated temperatures of Eocene and Oligo-Miocene magmatic rocks from the AMB ranges from 760 °C to 860 °C and 700 °C to 950 °C, respectively. These ages confirm temperatures obtained for the magmatic rocks by TZircsat.

Using the composition of the Lale-Zar zircons from Oligocene intrusions of SE UDMB, the fO_2 during zircon crystallization ranges between FMQ −0.69 to +2.41, whereas those of Hashroud, Gormolla and Techmedash zircons from Eocene, Oligocene and Miocene intrusions of NW AMB range from −0.20 to +2.6 (mean of +0.76), +1.9 to +5.7 (mean of +3.4) and + 1.6 to +4.3 (mean of +3.3) (Table 1; Supplementary Table 4c; Fig. 9a).

Crustal thickness of the UDMB and AMB during Cenozoic time

Calculated crustal thickness for the late-Cretaceous to late-Eocene, Oligocene and early-Miocene to late-Miocene magmatic rocks from the UDMB using Eq. (7) range from ~17 to 59 km (mean ~ 30 km), from ~12 to 49 km (mean ~ 27 km), from 11 to 58 (mean ~ 27 km), and from 70 to 170 km (mean ~ 112 km), respectively (Table 1; Supplementary Table 1a). Calculated crustal thickness of for the late-Cretaceous to late-Eocene, Oligocene–and early-Miocene to late-Miocene magmatic rocks from the UDMB using Eq. (8) range from ~6.0 to 40 km (mean ~ 22 km), from 8 to 94 km (mean ~ 32 km), from 23 to 53 (mean ~ 32 km), and from 56 to 107 km (mean ~ 72 km), respectively (Table 1; Supplementary Table S1a).

The calculated crustal thickness for the Eocene, Oligocene and early- to Late-Miocene of magmatic rocks from AMB using (Eq. (7)) range from ~14 to 57 km (mean ~ 32 km), from 24 to 59 km (mean ~ 38 km), and from 33 to 84 km (mean ~ 54 km) for the, respectively. The Eq. (5) for the same rocks range from ~24 to 59 km (mean ~ 38 km), from 45 to 87 km (mean

~ 60 km), and from 25 to 93 km (mean ~ 63 km), respectively (Table 1; Supplementary Table S1b).

6. Discussion

The thickness of the continental crust varies from just a few kilometers thicker than oceanic crust in island arcs (e.g. ~10 km) to over 80 km at some convergent margins, such as in the Himalaya–Tibet Orogen (Laske et al., 2013). Increasing crustal thickness in an arc setting can affect trace and major elemental chemistry of magmatic products (Chapman et al., 2015; Chiaradia, 2015; Profeta et al., 2015), and is also correlated with increased crustal assimilation (Haschke et al., 2002), quantified via oxygen fugacity. These features allow distinguishing between magmas from different geodynamic settings (e.g, mid-ocean ridges, intraplate). Typically, earc magmas show high Sr/Y values (e.g., Chiaradia, 2015). As such, we statistically analyzed whole-rock and zircon geochemical data from the UDMB and AMB regions to: (i) interpret magmatic processes and crustal thicknesses by whole-rock compositions; (ii) decipher magmatic condition by using zircon chemistry; and (iii) derive implications for the nature of Cenozoic arc magmatism.

5.1. Magmatic processes and thickness of crust inferred from whole-rock chemistry

The most of UDMB and AMB igneous rocks have calc-alkaline to shoshonitic signatures and show geochemical characteristics of I–type granitic rocks. Geochemical signatures of the igneous rocks indicate arc magmatism, including enrichment in LREEs (e.g., La, Ce), and high Ba/La (3.0–187) and low Nb/La (0.17–1.9) ratios compared to N-MORB (Sun and McDonough, 1989). Overall, the intermediate to felsic rocks of the UDMB and AMB magmatic rocks have high contents of Zr, Hf and Y, typical of continental-arc magmas. Although these geochemical characteristics imply a crustal contribution, due to lack of S-type granite, crustal anataxis of supracrustal sequences cannot be ruled in the petrogenesis. However, it is isotopically and geochemically very difficult to distinguish between mantle melt and melt evolving from the partial melting of the juvenile lower crust. Chaharlang et al. (2020) reported that initial εNd of UDMA magmatic rock fall within a range of of −5 to +5, resulting of interaction between juvenile mantle-derived melts (εNd ~ +8) and less radiogenic Cadomian continental crust of Iran (average εNd ~ −6). They also revealed a relationship between Nd isotope compositions and crustal thickness and suggested that the involvement of preexisting continental crust material increase by raising crustal thicknesses in the genesis of UDMB igneous rocks, based on the isotopic compositions.

The Sr/Y and (La/Yb)N ratios are often used to estimate the depths of magma crystallization (Eq. (7); Eq. (8)) (Chaharlang et al., 2020; Chiaradia, 2015). It has been suggested that Sr is preferentially incorporated into solid phases at low pressures (< ~1.0 GPa), where it strongly partitions into plagioclase during (i) the partial melting of lower crustal igneous rocks and their metamorphic equivalents, or (ii) during magmatic fractionation of mantle-derived mafic magmas (e.g., Profeta et al., 2015). Plagioclase becomes unstable at higher pressures (>1.2 GPa) in the mafic and intermediate compositional systems. Under such pressures, Sr favorably partitions into the melt phase, whereas Y is incompatible at low pressures, but readily partitions into garnet at high pressures (e.g., Profeta et al., 2015). Therefore, a high Sr/Y ratio of arc-related igneous rocks indicates crystallization at pressures (or depths) below the Moho (Fig. 9b), as is reported for the distinctive concentrations in the residual sub-arc rocks (e.g., Ducea, 2002). A Sr/Y ratio of more than 60 is expected if a melt is formed by partial melting of a subducted oceanic slab, whereas a Sr/Y of 30–50 is more representative of thickened continental crust, due to the greater Sr contents in altered mid-ocean ridge basalt relative to sub arc basalts/gabbros (e.g., Otamendi, 2009) (Fig. 9c).

The positive relationship between Sr/Y and (La/Yb)N ratios, and crustal thickness (Fig. 9b and c) in almost all the Cenozoic igneous rocks within Iran strongly argues for most of them having formed due to partial melting or crystal fractionation in the crust (Castillo, 2012), with slab melting deemed to have a minor to negligible role.

Most of the UDMB and AMB intrusions have Sr/Y ratios matching global Sr/Y values for crustal thicknesses exceeding ~22 km, and potentially to more than 30 km, as reported by Chiaradia (2015) (Fig. 9b and c). These values aresimilar to the estimations of crustal thicknesses ranging between 24.1 and 36.7 km based on (La/Yb)N ratios (3.1–5.5). A caveat for this type of analysis is that the crustal thickness achieved may be underestimated due to complex transitions across the Moho, involving mantle-like cumulates, such as gabbronorites to evolving pyroxenites, with or without garnet, instead of an abrupt change from typical lower crustal rocks to a peridotite. Consequently, the Iranian arc crust may have been thicker than predicted by these estimates.

Fig. 9b and c show Sr/Y and (La/Yb)N ratios as a function of time for the Cenozoic magmatic arcs of the UDMB and AMB, whose tectonic evolution is individually constrained by various geologic data. This is classically related to an across-arc spatial average and a temporal average discretized to ~100 Myr periods of time, while the numbers differ from region to region, depending on local particularities of magmatism and data availability (Profeta et al., 2015). The results show that the more recent crustal thickening began during the Late Cenozoic orogeny (10 to 5 Ma) in the UDMB, although it initiated earlier (40 to 10 Ma) in the northwest UDMB or northwest AMB. Magmatic bodies in the Cenozoic central UDMB have a relatively constant estimated crustal thickness from 50 to 10 Ma, and so does not appear to have thickened to the same extent as adjacent regions (Fig. 10)..

5.2. Magmatic processes inferred from zircon trace element composition

The most of UDMB and AMB igneous rocks have calc-alkaline to shoshonitic signatures and show geochemical characteristics of I–type granitic rocks. Geochemical signatures of the igneous rocks indicate arc magmatism, including enrichment in LREEs (e.g., La, Ce), and high Ba/La (3.0–187) and low Nb/La (0.17–1.9) ratios compared to N-MORB (Sun and McDonough, 1989). Overall, the intermediate to felsic rocks of the UDMB and AMB magmatic rocks have high contents of Zr, Hf and Y, typical of continental-arc magmas. Although these geochemical characteristics imply a crustal contribution, due to lack of S-type granite, crustal anataxis of supracrustal sequences cannot be ruled in the petrogenesis. However, it is isotopically and geochemically very difficult to distinguish between mantle melt and melt evolving from the partial melting of the juvenile lower crust. Chaharlang et al. (2020) reported that initial εNd of UDMA magmatic rock fall within a range of of −5 to +5, resulting of interaction between juvenile mantle-derived melts (εNd ~ +8) and less radiogenic Cadomian continental crust of Iran (average εNd ~ −6). They also revealed a relationship between Nd isotope compositions and crustal thickness and suggested that the involvement of preexisting continental crust material increase by raising crustal thickenesses in the genesis of UDMB igneous rocks, based on the isotopic compositions.

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Fig. 11 present the linear correlations between Sr/Y values at intermediate MgO (2–6 wt%) contents which show that variable Sr/Y values of the most recent arc magmas have been controlled by the upper plate thickness. This interpretation suggests that slab melting or slab melt-mantle interactions did not play a main role in the formation of high Sr/Y signatures. Chiaradia (2015) proposed that the average of Sr/Y values of less than 20-km-thick arcs remains constant (~15 km) where MgO values vary from 6 to 10 wt%, and then gradually decrease where MgO contents fall below 6 wt% (Fig. 11a). Our compiled data from the UDMB and AMB units also show a similar trend with global Sr/Y correlation with crustal thickness of Chiaradia (2015), suggesting magmatic fractionation in the stability field of plagioclase (Fig. 11 b,c). The Sr/Y ratios of rocks versus Moho depth situated at >30 km steadily increase during magmatic differentiation (Fig. 11d), starting with Sr/Y ratios of 20 at 10 wt% MgO and rising to Sr/Y ratios of 40–50 at intermediate-low MgO values (2–4 wt%), proposing magmatic fractionation in the stability field of plagioclase. The UDMB and AMB rocks recording crustal thicknesses at >30 km also display similar trend with global Sr/Y correlation with crustal thickness of Chiaradia (2015) (Fig. 11 e, f). The UDMB and AMB rocks recording crustal thicknesses of 20–30 km show also similar trend between Sr/Y and MgO, which plots between those of the two end-members discussed above (Fig. 11g-i). The Sr/Y ratios of all rock types with 10 wt% MgO increase with crustal thickness (Fig. 11). Moreover, the northwest UDMB and AMB samples show greater crustal thickness relative to the central and southeast UDMB and central and northeast AMB (Fig. 11).

We consider the ranges of $(Sm/Yb)_N$ and the Eu anomaly (Eu/Eu*) of the magmatic rocks from UDMB and AMB with respect to their thickness. A $(Sm/Yb)_N$ below 4 or above 5, reflects amphibole- or garnet-dominated source of fractionated granitoids, respectively.

Most of the crustal thicknesses of less than 20 km and 20–30 km calculated based on the whole rock geochemistry of AMB correspond to $(Sm/Yb)_N < 4$, while those of more than 30 km have values up to $(Sm/Yb)_N > 4$ (Supplementary Table S1b). However, the UDMB samples display a somewhat similar $(Sm/Yb)_N < 4$ for over-thickened crust (Supplementary Table S1a), suggesting that amphibole had a major role in fractionating the REE, perhaps in the presence of garnet, but certainly not dominated by garnet.

Magmatic processes inferred from zircon trace element composition

Trace element data of the UDMB and AMB magmatic zircons define a single group in terms of their REE contents and chondrite-normalized REE. The LREE contents are moderately low among the values reported for crustal zircons (~250–5000 ppm; Hoskin and Schaltegger, 2003) ranging from 100 to 2500 ppm for both the UDMB and AMB (Suppementary Table S1c). Most of the studied zircons from the southeast and northwest UDMB/AMB show Ti in zircon temperatures (705–920 °C) and U/Th ratios corresponding to igneous origin (Siégel et al., 2017). This is in agreement with experimental data on zircon temperature saturation in magmas that produce intermediate rocks, such as those formed in Phanerozoic subduction-related magmatic arcs (Watson et al., 2006). Thus, analyses of these zircon populations allows a first-order interpretation of the magmatic processes involved in their formation. However, it should be mentioned that the LREE contents up to several thousand ppm are excluded since they have been modified by secondary processes and are not representative of the magmatic processes. Compiled data from the UDMB and AMB intrusions show an increase in average zircon crystallization temperatures with decreasing age (Table 1; Supplementary Table S4).

A positive Ce anomaly in chondrite-normalized REE patterns of zircon from the UDMB and AMB magmatic rocks indicates that Ce^{4+} was stable in the magma when the zircon crystallized. In addition, all zircons display negative Eu anomalies (Jahangiri, 2007 Kananian et al., 2014; Sarjoughian and Kananian, 2017), which implies either (i) the zircons crystallization under reducing conditions; or (ii) plagioclase is formed before crystallization of zircon, since Eu was preferentially incorporated within plagioclase. Nonetheless, the negative Eu anomalies of the studied zircons could have been caused by a combination of these processes, and do not directly reflect the oxygen fugacity when zircons are formed. Burnham and Berry (2012) reported a correlation between positive Ce and negative Eu anomalies in zircon and the oxygen fugacity at which it was formed. Zircon does not acquire a positive Ce anomaly when $\log fO_2$ is lower than about −10. These results argue that the zircons from the UDMB and AMB samples could not have occurred at $\log fO_2 < -10$ (Fig. 9).

Implications for crustal thickness and magma evolution of the Cenozoic magmatic belt

The high Sr/Y (>56) and La/Yb (>20) ratios of the Plio-Quaternary rocks point to their fractionation from a garnet-, amphibole- and clinopyroxene-bearing source (e.g., Castillo, 2012; Hernández-Montenegro et al., 2021). The adakitic rocks of the UDMB show high La/Sm (7.3–10.6) and Sm/Yb (2.3–7.6) ratios that are similar to that of Andean-type arc (~60 km thickness) magmas (Fig. 12d), signifying a LREE-enriched source and/or crustal assimilation or a garnet-bearing metasomatized source underneath the continental arc (Rapp et al., 1999).

In contrast, the Eocene-Oligo-Miocene (60 to 15 Ma) and older igneous rocks of the UDMB show a non-adakitic affinity and their Sr/Y (<21) and La/Yb (<9) ratios are identical to island-arc magmatism (Fig. 12a-c). They are characterized by low La/Sm (1.8–4.2) and Sm/Yb (0.9–3.8) ratios, suggesting a less LREE-enriched and pyroxene- and plagioclase-dominated residual mineralogy typical of a thin island arc (~30–35 km thickness; Mantle and Collins, 2008) (Fig. 12d) compared to thick sub arc crust.

The Cenozoic igneous rocks of the AMB have non-adakitic signatures with Sr/Y (<21) and La/Yb (<9) ratios that are akin to those of island-arc magmatism (Fig. 13a,b). Some of the igneous rocks with ages of 30 to 5 Ma display more adakitic signatures in a plot of (La/Yb vs. SiO_2 (Fig. 13c). It is believed that these are intrusions that were generated from mantle sources (Castro et al., 2013). However, we suggest that they could be sensitive for the crustal evolution of the AMB, as the intrusions crystallized at the depths of crustal levels. The

Cenozoic igneous rocks of the AMB exhibit a similar trend, which relates to an increasing crustal thickness with decreasing age, and widely variable La/Sm and Sm/Yb ratios from an Andean-type arc (~60 km thickness) scenario (Fig. 13d) to 45 km. This indicates that a LREE-enriched source is responsible for the generation of the younger rocks (e.g., Rapp et al., 1999).

Some of the magmatic rocks of the AMB at 30 to 5 Ma also display adakitic signatures with high Sr/Y (>56) and La/Yb (>20) ratios (Fig. 13a and b), indicating garnet, hornblende, and clinopyroxene minerals in the source (Castillo, 2012). These adakitic rocks show high La/Sm (7.3–10.6) and Sm/Yb (2.3–7.6) ratios that are similar to those of the Andean-type arc (~60 km thickness) (Fig. 13d). This interpretation also points to a LREE-enriched source (e.g., or a garnet-bearing metasomatized source underneath a continental arc (Rapp et al., 1999).

There is still considerable debate on the origin of adakitic rocks that can be generated via several processes: (i) high- and low-pressure fractionation (Gao et al., 2009) of hydrous basaltic magma, (ii) partial melting of a subducting oceanic slab (Martin, 1999), (iii) mixing between mafic and felsic magmas (Qin et al., 2010), (iv) low-degree partial melting of metasomatized mantle (Gao et al., 2009), and (v) melting of a thickened mafic lower crust (Karsli et al., 2010). Different geochemical features of adakitic rocks are expected when they are generated in different geological environments (Hollings et al., 2005).

In the La/Sm vs La diagram (Fig. 14 a,c), most of the Cenozoic rocks follow a partial melting trend rather than a fractional crystallization trend. A linear trend in the AMB magmatic rocks suggests that the igneous rocks could not have been produced by a low-pressure fractional crystallization process involving a parental basaltic magma. It can be seen that magmatic rocks with older ages have lower partial melting degrees than younger rocks (Fig. 14a, c). As mentioned above, igneous rocks with older ages (Early Cenozoic) correspond to lower crustal thickness, whereas Late Cenozoic rocks have greater crustal thickness signatures. A close relationship between adakitic rocks and crustal thickness of magmatic rocks from UDMB has also been proposed in previous works (e.g., Asadi et al., 2014), which suggested that most adakitic magmatism was related to melting of a thickened juvenile mafic lower crust. Therefore, we suggest that Early Cenozoic rocks of the UDMB experienced lesser assimilation and fractional crystallization (AFC) processes than the Early Cenozoic rocks of the UDMB and Late Cenozoic rocks of AMB (Fig. 14), which is in agreement with crustal thickness of both the groups (Fig. 9b and c).

The batch-melting modeling on the diagram of (La/Yb) N vs. (Yb)N diagram (Drummond et al., 1996) shows that Paleocene to Miocene magmatic rocks of the UDMB were generated by ~3% melting of a presumed garnet amphibolite source (Tepper et al., 1993), with minor plagioclase as a residual mineral (Fig. 14c). This agrees with the low degrees of partial melting for these rocks predicted via La vs. La/Sm characteristics (Fig. 14a). The late-Miocene to Pliocene rocks likely experienced ~10–15% (mostly >10%) melting of a garnet and amphibolite bearing source (Tepper et al., 1993), with plagioclase being a minor residual mineral (Fig. 14a and c; e.g., Karsli et al., 2011).

The rocks from Paleocene to Pliocene are formed by 3–10% partial melting of a presumed amphibolite/garnet amphibolite source and that some rocks included in that are formed by up to 30% (Fig. 14d); however, some of the late Miocene and Pliocene rocks exhibit higher degree partial melting (~15–30% of a garnet amphibolite source; Fig. 12d). The average (La/Yb)N of the lower continental crust is about 5.3 (Rudnick and Gao, 2003), which is over six times higher than that of average MORB (~0.8; Sun and McDonough, 1989). Therefore, magmas formed by partial melting of the lower continental crust in the presence of garnet should have systematically higher (La/Yb)N than those of slab melts (Fig. 14), regardless of the amount of plagioclase present in the source (Sun et al., 2012). The batch-melting modeling indicates that melting of a garnet-free basaltic amphibolite source can explain the formation of most of the Paleocene to Early Miocene rocks in the studied regions. Moderate degrees of amphibolite melting (5–20%) and subsequent fractional crystallization of plagioclase and pyroxene are consistent with the low (La/Yb)N ratios (<6.7) of the these magmatic rocks (e.g., Karsli et al., 2011) (Fig. 14).

Collision of the Arabian and Iranian plates in the Late Cenozoic (Gammons and Williams-Jones, 1997) resulted in the thickening of the juvenile arc mafic lower crust and metamorphism to garnet amphibolite (Cooke et al., 2005) (Fig. 15a). Gravimetric and whole-rock geochemistry data show that crustal thickness increased due to the collision, and the thickness reached ~50–60 km during the Neogene (Dehghani and Makris, 1983) (Fig. 15b), and the garnet amphibolite crustal root with thickened lithospheric mantle became unstable and began to delaminate (e.g., Karsli et al., 2010). The lithospheric mantle thinning led to trans-tensional faulting, such as the Rafsanjan fault, situated to the southeast of the UDMB, and formed adakitic magmas.

7. Conclusions

Sr/Y and (La/Yb)N ratios as well as intensive parameters such as fO2 and temperature for the Eocene to middle Miocene igneous rocks of the UDMA and AMB display the following information concerning historical variations in the crustal thickening during the Cenozoic:

The Late Cenozoic magmatism of the UDMB and AMB shows a significant pulse of Sr/Y and La/Yb increase, separated by period of higher Sr/Y and La/Yb.

The oxygen fugacity (fO2) of Lale-Zar zircons from the SE UDMB is characterized by FMQ of −0.69 to +2.41, during zircon crystallization, whereas those of the Hashroud-Techmdash-Gormolla from NW UDMB range from −1.22 to +5.99. The fO2 estimates suggest relatively more oxidized conditions for the Late Cenozoic igneous rocks of the Urumieh-Dokhtar and Alborz magmatic belt.

Compiled data from the UDMB and AMB intrusions show an increase in average zircon crystallization temperatures with decreasing age.

These outcomes have been interpreted in terms of variation of the crustal thickness, from 30 to 35 km during Eocene-Oligocene to 40–55 km during the middle–late Miocene. Here, we suggest that the increase in the crustal thickness resulted from the collision between the Arabian and Iran plates and subsequent convergence during the middle–late Miocene.

Declaration of Competing Interest

We declare that we have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Figure captions

Figure 1) Map of Iran showing the Cadomian basement rocks, Cenozoic igneous rocks and Paleozoic–Mesozoic ophiolites.

Figure 2) Schematic stratigraphic column displaying idealized internal lithologic succession in the (a) UDMB and (b) AMB.

Figure 3) (a) and (b) Total alkalis versus SiO_2 diagram for classification of the igneous rocks from UDMB and AMB rocks (after LeBas et al. (1986)); (c) and (d) A/NK *vs*. A/CNK diagram for the studied rocks; (e) and (f) K_2O *vs*. SiO_2 diagram with field boundaries between medium-K, high-K and shoshonitic series according to Peccerillo and Taylor (1975). Field boundaries between I-type and S-type granitoids are from Chappell and White (1974), and peraluminous and metaluminous fields are from Shand (1943). References for location are mentioned in text.

Figure 4) (a) and (b) Chondrite-normalized rare earth element of AMB (left) and UDMB (right). Chondrite and N-MORB normalized values are taken from Sun and McDonough (1989).

Figure 5) Zircon trace elements plots from the UDMB and AMB igneous rocks (chondrite normalized values are taken from Sun and McDonough, 1989).

Figure 6) Weighted mean age diagrams for UDMB and AMB magmatic rocks.

Figure 7) Whole-rock (a) La/Yb and (b) Sr/Y correlations with crustal thickness for modern subduction-related volcanic arcs. La/Yb ratios are normalized to chondritic values.

Figure 8) Average chondrite-normalized La/Yb and Sr/Y versus time (million years) for UDMB and AMB.

Figure 9) Sr/Y versus MgO plots of magmatic rocks from UDMB Cenozoic arc. Small grey dots are individual rock analyses from the Georoc database with their density distribution (points/cell) indicated by color shadings. The large black symbols (circles N, squares &, and triangles m) are averages of median MgO and Sr/Y values for intervals equal or bigger than 0.5 wt.% MgO for arcs, 20 km, 30 km, and 20–30 km thick, respectively. Bars are 1 sigma associated with the averages of median values (bars for MgO values are often smaller than symbol size).

Figure 10) Sr/Y versus MgO plots of magmatic rocks from AMB Cenozoic arc. Small grey dots are individual rock analyses from the Georoc database with their density distribution (points/cell) indicated by color shadings. The large black symbols (circles N, squares &, and triangles m) are averages of median MgO and Sr/Y values for intervals equal or bigger than 0.5 wt.% MgO for arcs 20 km, 30 km, and 20 to 30 km thick, respectively. Bars are 1 sigma associated with the averages of median values (bars for MgO values are often smaller than s

Figure 11**.** Values of log*f*O_2 calculated using the Ce-in-zircon oxygen barometer presented in this study as a function of temperature for zircons from the UDMB and AMB. Independent estimates of *f*O_2 and *T* for the UDMB and AMB samples are derived from Fe–Ti oxide (Chesner, 1998; Hildreth and Wilson, 2007). Typical error in calculated *f*O_2 is shown in the upper left corner (approximately ±1 log unit). The dashed black curves correspond to the variation in *f*O_2 with *T* along the FMQ buffer, calculated from O'Neill (1987), the magnetite–hematite (MH) and iron–wustite (IW) buffers calculated from O'Neill (1988).

Figure 12 Diagrams of (a) Sr/Y vs. Y and (b) La/Yb vs. Yb (from Defant and Drummond, 1993), black arrow trend indicate the presence of hornblende, clinopyroxene, titanomagnetite and plagioclase in the source area (after Castillo, 2012); (c) La/Yb vs. SiO2 (after Richards et al., 2012) (d) Plots of Sm/Yb vs. La/Sm ratios for UDMB igneous rocks (PYX = pyroxene, AMPH = amphibole, GAR = garnet; approximate boundaries from Kay and Mpodozis (2001) of KCMA porphyry rocks in comparison with Andean orogenic arcs (Cordillera Blanca batholith, Peru; data from Petford and Atherton, 1996; El Abra-Fortuna batholith at Chuquicamata Cu mine, Chile; data from Haschke et al., 2002; Haschke et al., 2010).

Figure 13 Diagrams of (a) Sr/Y vs. Y and (b) La/Yb vs. Yb (from Defant and Drummond, 1993), black arrow trend indicate the presence of hornblende, clinopyroxene, titanomagnetite and plagioclase in the source area (after Castillo, 2012); (c) La/Yb vs. SiO2 (after Richards et al., 2012) (d) Plots of Sm/Yb vs. La/Sm ratios for AMB igneous rocks (PYX= pyroxene, AMPH = amphibole, GAR = garnet; approximate boundaries from Kay and Mpodozis (2001) of KCMA porphyry rocks in comparison with Andean orogenic arcs (Cordillera Blanca batholith, Peru; data from Petford and Atherton, 1996; El Abra-Fortuna batholith at Chuquicamata Cu mine, Chile; data from Haschke et al., 2002; Haschke et al., 2010).

Figure 14. (a) La/Sm vs. La (ppm) and (b) batch-melting modeling of chondrite-normalized [La/Yb]n ratios vs. [Yb]n diagrams after Drummond et al. (1996) for UDMB igneous rocks. (c) La/Sm vs. La (ppm) and (b) batch-melting modeling of chondrite-normalized [La/Yb]n ratios vs. [Yb]n diagrams after Drummond et al. (1996) for AMB igneous rocks. Fields of adakite-like rocks and island-arc type are from Martin (1999). An Eastern Pontides gabbro (G518) (Dokuz et al., 2006) is used as the source rock for the REE modeling under amphibolite and eclogite conditions, with varying garnet contents and respective partition coefficients (I–VI) proposed by Irving and Frey (1978), Fujimaki et al. (1984), and Sisson (1994). Data for delaminated lower crust-derived adakitic rocks are from Xu et al. (2002) and Wang et al. (2004). The field of subducted oceanic crust-derived adakites is constructed using data from the following: Defant and Drummond (1990), Kay et al. (1993), Drummond et al. (1996), Stern and Kilian (1996), Sajona et al. (2000), Aguillón-Robles et al. (2001), Defant et al. (2002) and Martin et al. (2005), and references therein. Data for thick lower crust-derived adakitic rocks are from the following: Atherton and Petford (1993), Muir et al. (1995), Petford and Atherton (1996), Johnson et al. (1997) and Xiong et al. (2003). Pure slab melts are after Kepezhinskas et al. (1995) and Sorensen and Grossman (1989). Other diagram for the KCMA porphyries, indicating Kuh Panj evolution is mainly controlled by partial melting.

Figure 15. Simplified cross-section of the (a) Early to Middle Cenozoic and (b) Late Cenozoic magmatism in Iran, showing positions of the UDMB and AMB; subcontinental lithospheric mantle (SCLM) melting and assimilation/fractional crystallization (AFC) or melting-assimilation-storage-homogenization (MASH) processes. in AMB vs UDMB in Early to Middle Cenozoic magmatism.