

Petrogenesis of Tarom high-potassic granitoids in the Alborz–Azarbaijan belt, Iran: Geochemical, U–Pb zircon and Sr–Nd–Pb isotopic constraints



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ABSTRACT

Large-scale Upper Eocene plutons in the Western Alborz–Azarbaijan orogenic belt mostly show calc-alkaline and I-type geochemical features contrasted by the Tarom complex with its high-potassic to shoshonitic affinity. The pluton was emplaced in the Tarom subzone of the orogenic belt and its laser ICP-MS zircon U–Pb age of 41 Ma is interpreted as the age of magma crystallization. The Tarom complex is composed of quartz monzodiorite, quartz-monzonite and monzogranite, the SiO₂ contents range from 57 to 70 wt.%, the K₂O + Na₂O content is high (5.0–8.9 wt.%) and K₂O/Na₂O ratio ranges from 0.4 to 1.9. All the investigated rocks are enriched in light rare earth elements (LREEs), large ion lithophile elements (LILEs) and depleted in high field strength elements (HFSEs), and bear a weak Eu anomaly (Eu/Eu* = 0.46 to 1.38) in chondrite-normalized trace element patterns. The samples display some variety in initial Sr and Nd isotopic compositions, marked with low I_{Sr} = 0.704–0.705 and ε_{Nd} (40 Ma) = −4.2 to +3.4 (−5.7 for an enclave) values. The Pb isotopic ratios are (²⁰⁶Pb/²⁰⁴Pb) = 18.52–18.86, (²⁰⁷Pb/²⁰⁴Pb) = 15.57–15.72 and (²⁰⁸Pb/²⁰⁴Pb) = 38.47–39.08. Comparison with experimental studies, together with mantle-like isotopic ratios and comparisons of REE patterns, points to an origin of chemically enriched lithospheric mantle source for the Tarom plutonic complex. Partial melting process involving different partial melting degrees affecting heterogeneously metasomatized mantle is a process that seems likely to have occurred in the studied complex as the major differentiation process. The Tarom monzonitic plutons are considered to be post-orogenic intrusions that were emplaced in an environment of lithospheric extension, causing asthenospheric upwelling. Asthenospheric upwelling induced a thermal anomaly which caused partial melting of metasomatized subcontinental lithospheric mantle in the Tarom area.

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1. Introduction

In general, shoshonitic magmas are assumed to be transitional between calc-alkaline and alkaline magmas (Bonin, 2004; Liewe et al., 1998). Although many high-potassic plutons have been studied in detail, their generation and geodynamic settings are still under considerable debate. High potassium and shoshonitic magmatism is characteristically appearing in both collisional and subduction-related orogenic environments. Muller et al. (1992) also linked the occurrence of high-potassic rocks to five different tectonic settings: continental arcs, post-collisional arcs, oceanic arcs (subdivided into initial and late), and within-plate, trying to discriminate them with geochemical parameters.

However, Turner et al. (1996) revealed that post-collisional, potassic magmatism is a common feature in many collisional orogenic belts around the world. There are several references for high potassic and shoshonitic granitoids which suggest post-collisional arc setting for generation of this type of magmatism (e.g., Aghazadeh et al., 2011; De Lima and Nardi, 1998; Karsli et al., 2007, 2012; Lan et al., 2012; Wang et al., 2007).

Shoshonitic magmas may form: (i) by partial melting of enriched mantle without assimilation of crustal material during ascent (e.g., Jiang et al., 2002; Liu et al., 2002; Wang et al., 2006), (ii) by partial melting of enriched mantle with assimilation of crustal material (e.g., Duchesne et al., 1998; López-Moro and López-Plaza, 2004), and (iii) by partial fusion of depleted mantle with subsequent contamination by crustal material (e.g., Fu et al., 1997; Xing and Xu, 1999) in both arc and post-collisional settings (e.g., Foley and Peccerillo, 1992; Turner et al., 1996). Because an enriched mantle source seems to be involved in the generation of high-potassic and shoshonitic magmas, the implication

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of subduction components is straightforward. Monzonitic magmas can be generated as primary melts by decompression melting (1.0–1.5 GPa) of a metasomatized lithospheric mantle within the stability field of pargasite (Conceição and Green, 2004). A subduction-related metasomatized mantle source is also proposed for the high potassic eruptive magmas in the Oligocene–Pleistocene western Mediterranean region (Conticelli et al., 2009).

New petrological data from the high-K calc-alkaline Tarom complex in the northwest of the Zagros–Alborz orogenic system are reported here for the first time with the aim of contributing to the better understanding of the tectonic evolution of this important sector of the Tethyan orogen.

2. Regional setting of Alborz Mountains

Iran and the surrounding areas present a mosaic of continental blocks separated from each other by complex fold–thrust belts within the Alpine–Himalayan orogenic system (Gansser, 1981). The oldest basement of the Iranian plate is located in the Central Iranian Terrane (Fig. 1) and is composed of a Precambrian basement with an Infracambrian to Mesozoic cover (e.g., Förster and Jafarzadeh, 1994). Most Neoproterozoic paleogeographic reconstructions (e.g., Stampfli and Borel, 2004) and paleomagnetic studies (Becker et al., 1973; Soffel and Förster, 1981) place the Central Iranian Terrane alongside the Arabian and Indian plates, along the Proto-Tethyan margin of the Neoproterozoic–Early Cambrian Gondwanaland.

Following the formation of the Proto-Tethys, Iran was controlled by the opening and closure of the Tethyan oceans of different successive stages. The idea that two successive and partly contemporaneous

Tethyan oceans existed, the northerly, older Paleo-Tethys and the southerly, younger Neo-Tethys, has currently become a consensus among workers (e.g., Stampfli, 2000). In general, the Tethyan orogen formed from the collision of Eurasia with dispersed fragments of Gondwanaland (Sengör, 1984; Sengör and Natal'in, 1996).

According to regional tectonics, magmatic sequences, paleogeography and ophiolite remnants, the Paleo-Tethys Ocean separated the Iranian plate from Eurasia during the Late Paleozoic–Early Mesozoic times and the Neo-Tethys Ocean was located between the Iranian and Arabian plates during the Late Mesozoic times (Stocklin, 1974). Subduction of the Paleo- and Neo-Tethys beneath the Eurasian and Iranian plates, respectively, caused the formation of the Caucasus magmatic arc in the north and the Sanandaj–Sirjan Magmatic Zone, the Urumieh–Dokhtar Magmatic Arc, the Alborz Mountain belt and the Zagros fold-and-thrust belt in the south.

The Alborz magmatic belt is situated in northern Iran and has an E–W orientation 600 km in length and 100 km wide and is separated from the Central Iranian microplate in the south by the Tabriz Fault (Fig. 1); to the north, the belt extends into Armenia (Azizi and Jahangiri, 2008). This belt is divided into western and eastern parts by the N–S trending Rasht–Takestan Fault. The Alborz magmatic belt occurs in the hinterland of the Arabia–Eurasia collision zone in the broad Alpine–Himalayan orogenic belt.

According to Sengör et al. (1993) and Zanchi et al. (2006), the Alborz range results from different tectonic events, which occurred during two major orogenic cycles: (1) the Late Triassic (Cimmerian) demise of the Paleo-Tethys and collision of the Iranian block with Eurasia; and (2) the post-Neo-Tethys, intraplate deformation related to the convergence of the Arabian and the Eurasian plates. In the Alborz magmatic

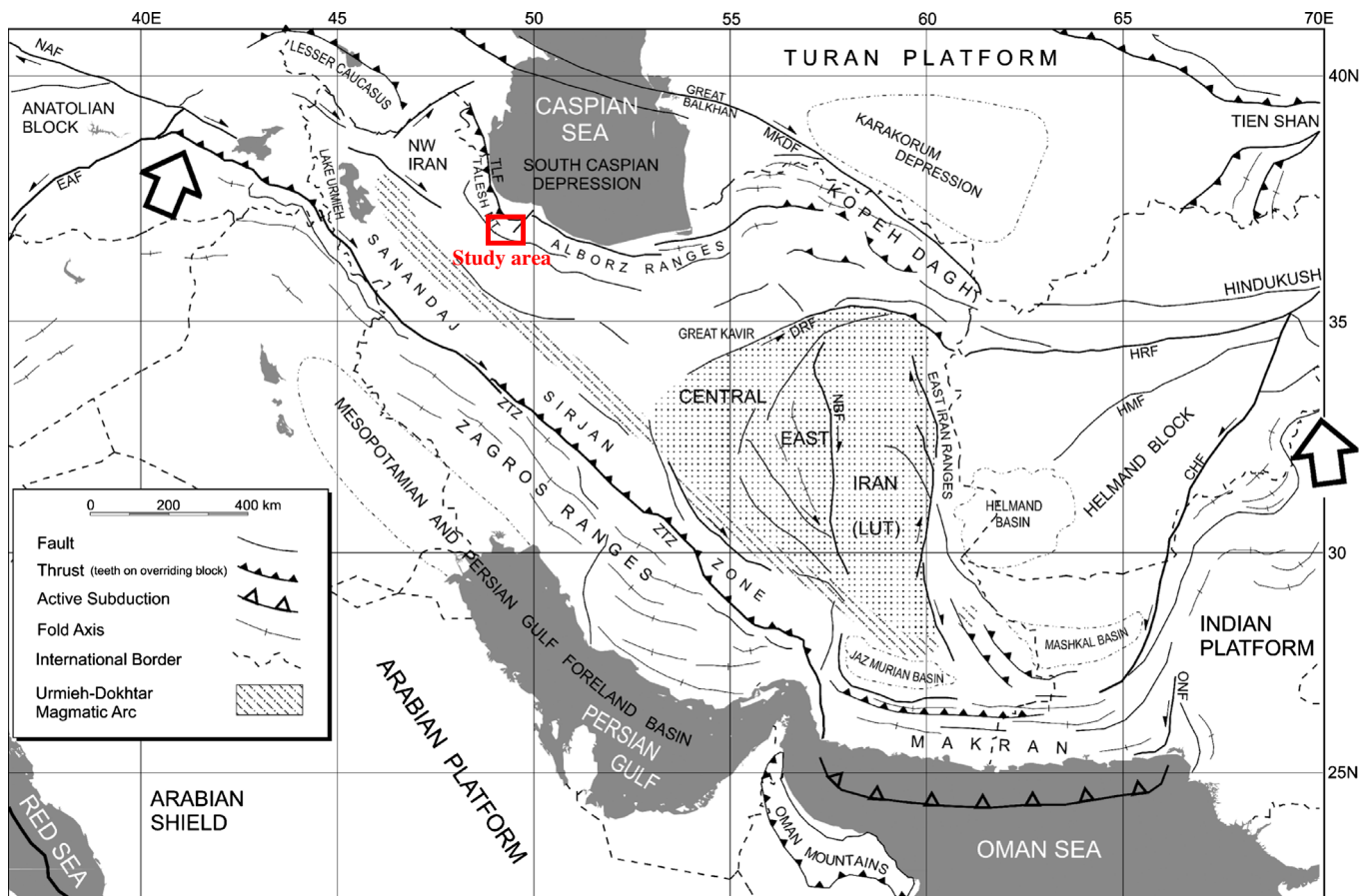


Fig. 1. Simplified structural map of Iran and adjacent regions and location of Tethyan sutures. Large arrows illustrate the syntaxes of the Alpine–Himalayan Orogenic System. Modified after Ramezani and Tucker, 2003; actual motion of the Arabian plate relative to the Eurasia from Vernant et al. (2004).

belt, the bedrock consists of a Paleozoic–Mesozoic passive-margin sedimentary sequence with few volcanic rocks, overlain, in the south, by Tertiary volcanics and sediments (Mirnejad et al., 2010).

According to Bolourian (1994), the eastern part consists of basic and acidic tuff and lava with an alkaline to shoshonitic affinity with an Eocene age, while the western part (western Alborz–Azarbaijan magmatic belt) consists of Eocene andesitic to dacitic lava and many granitoid bodies with a calc-alkaline affinity (Moayyed, 2001; Nabatian and Ghaderi, 2013; Nabatian et al., 2013). These volcanic and plutonic rocks with calc-alkaline and alkaline (potassic) affinity are formed in arc and back-arc settings within an extensional (post-collisional) environment (Allen et al., 2003; Berberian, 1983; Hassanzadeh et al., 2002; Mirnejad et al., 2010).

Cenozoic successions of the Alborz magmatic belt are represented mainly by marine and subaerial, porphyritic and non-porphyritic, massive lava flows of andesite, basaltic andesite and basalt compositions (Aghazadeh et al., 2010). In the northwestern part of the Alborz

magmatic belt, Cenozoic strata were intruded by Upper Eocene and Oligo–Miocene plutons of shoshonitic and high-K calc-alkaline affinities, which are typically related to post-collisional magmatic activity (Aghazadeh et al., 2010; Berberian and King, 1981; Nabatian and Ghaderi, 2013; Nabatian et al., 2013, 2014; Verdel et al., 2011).

According to the data given above, there are some descriptions on Cenozoic magmatic products of the Alborz–Azarbaijan magmatic belt (e.g., Shaivar Dag, Khankandi, Anzan, Kalaibar, Sungun, and Ordobad; Aghazadeh et al., 2011; Babakhani et al., 1990; Berberian and Berberian, 1981; Mokhtari et al., 2010). However, no detailed studies on the genesis and tectonic setting of the plutonic complex have been conducted in the Tarom area.

3. Geological setting of the Tarom area

The study area is part of the Tarom volcano-plutonic complex of the Cenozoic Alborz–Azarbaijan magmatic belt (Fig. 2). Magmatism in the

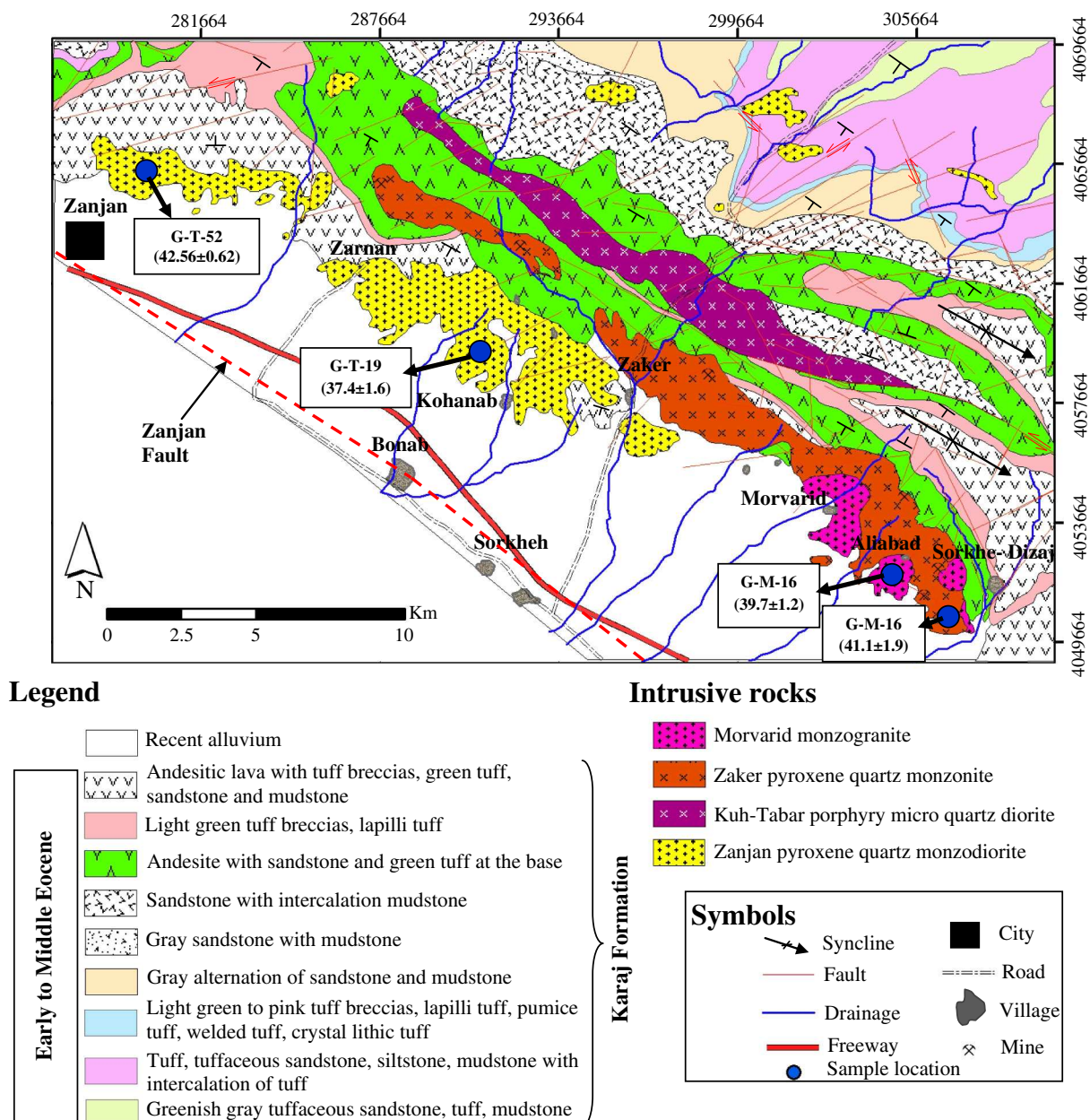


Fig. 2. Geological map of the Tarom intrusive complex. Simplified after Nabatian (2011).

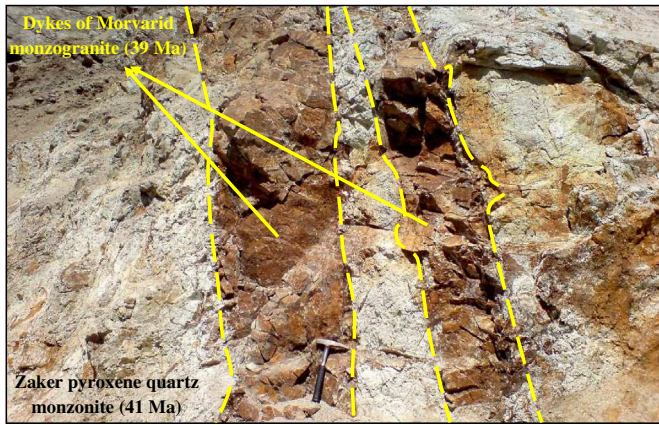


Fig. 3. Morvarid monzogranitic dyke cross-cutting the Zaker quartz monzonite.

area began during the Eocene and continued throughout the Late Eocene. This region is composed mostly of pyroclastic rocks (Karaj Formation; Fig. 2) later followed by the intrusion of late Eocene plutonic rocks. The Eocene marine volcanism includes pyroclastic and lava flows of trachyte, trachy-andesite, andesite, basaltic andesite, olivine-basalt, porphyritic and non-porphyritic rhyodacite with calc-alkaline to alkaline affinity (Bolourian, 1994; Moayyed, 2001; Nabatian, 2011). This period of intense volcanic activity is associated with sedimentary rocks such as limestone and sandstone at the base. Verdel et al. (2011) show that the zircon U–Pb ages from the Karaj Formation in the Alborz Mountains yielded 49.3 ± 2.9 Ma for the so-called Middle Tuff, 45.3 ± 2.3 Ma for a tuff within the Asara Shale and 41.1 ± 1.6 Ma for the Upper Tuff. According to Amini et al. (2001), the andesitic rocks, which host the Tarom plutonic complex, are equivalent to younger tuff units within the Karaj Formation.

Tarom plutonic complex is widespread throughout the Tarom sub-zone and forms a 20–25 km long elongated NW–SE trending body (Fig. 2). The plutonic bodies including Zanjan, Zaker, Morvarid (ZZM) and Kuh-Tabar intrusions (Fig. 2) have sharp contacts and include some mafic microgranular enclaves near the margins. Most of the mafic microgranular enclaves have quartz-diorite and quartz-monzonite compositions. Some monzogranitic dikes cross-cut the Zaker quartz monzonitic pluton (Fig. 3). The Kuh-Tabar pluton is elongated parallel to the ZZM, but there is no contact between the Kuh-Tabar and ZZM intrusive bodies. The Eocene andesites are metamorphosed near the contact with the plutonic units. The grade of contact metamorphism is locally low within the albite-epidote hornfels facies.

4. Petrography and mineral compositions

Tarom magmatic complex is composed of monzonitic intrusions and volcanic rocks. The monzonitic intrusions include Zanjan quartz-monzodiorite, Zaker quartz-monzonite, Morvarid monzogranite and Kuh-Tabar quartz-monzodiorite. These intrusions show some differences

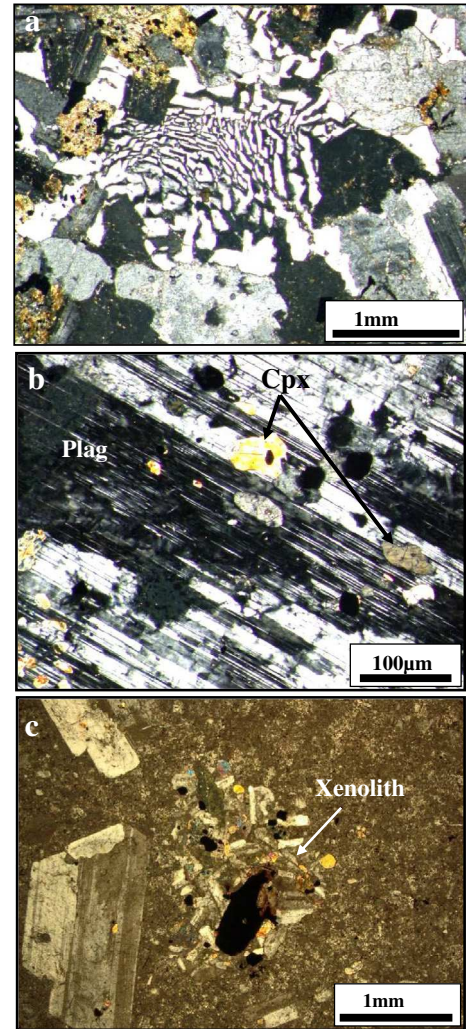


Fig. 4. (a) Granophyric textures in the Zaker plutonic rock. (b) Inclusions of pyroxene within the plagioclase in the Tarom complex. (c) Xenolith of Kuh-Tabar intrusive body within the Zaker plutonic rocks.

in texture and mineral assemblage. Here, petrographic and mineralogical characteristics of the main rock types distinguished in the Tarom area are described in detail.

4.1. Zanjan pluton

The pyroxene-quartz-monzodiorite is generally medium-grained and characterized by an equigranular texture, comprising plagioclase (60–64%), K-feldspar (25–27%), quartz (7–10%) and clinopyroxene (8–11%) (Table 1). Accessory minerals include zircon, magnetite, apatite

Table 1

Modal mineralogical composition of the Zanjan pyroxene-quartz-monzodiorite, Zaker pyroxene-quartz-monzonite and Morvarid monzogranite to syenogranite.

Sample	Zaker pyroxene-quartz-monzonite										Morvarid monzogranite to syenogranite				Zanjan pyroxene-quartz-monzodiorite				
	M 25	M 39	M 47–47	N-A1	N-SD19	GN1	G7	SD86	SD92	G20	M 60	N-SD4	N-M65	N-M16	N-T19	N-T38	N-T52	N-T57	N-Z8
Quartz	10	11	11	10	10	11	10	12	12	12	25	26	27	29	9	8	7	7	7
Plagioclase	41	41	39	43	42	40	40	38	39	40	27	24	25	23	58	59	64	60	60
K-Feldspar	38	37	38	36	37	40	39	40	41	39	43	46	45	45	25	24	23	24	25
Clinopyroxene	9	9	8	9	8	7	8	7	6	7	0	0	0	0	7	8	6	9	7
Biotite	0	0	1	0	0	0	0	0	0	0	4	3	2	2	0	0	0	0	0
Apatite	0.5	0.2	0.6	0.8	1.0	0.6	1.2	1.0	0.8	1.1	0.0	0.0	0.1	0.2	0.3	0.3	0.6	0.3	0.6
Opaque	1.1	0.9	1.7	0.8	1.75	1.3	1.5	1.7	1.1	0.8	0.4	0.3	0.4	0.4	1.1	1.2	0.7	0.8	0.8

Table 2
Microprobe analyses (wt.%) of plagioclase in the representative rock types of the Tarom plutons.

Sample	Zaker quartz monzonite				Kuh Tabar micro quartz monzodiorite				Zanjan quartz monzodiorite				Morvarid monzogranite	
SiO ₂	54.42	62.01	61.81	52.50	56.55	55.47	11.42	55.47	56.55	61.66	49.97	55.91	66.83	66.24
TiO ₂	0.09	0.03	0.01	0.07	0.05	0.04	0.05	0.04	0.05	0.05	0.01	0.04	0.13	0.00
Al ₂ O ₃	26.92	23.17	22.86	28.71	26.58	27.16	23.60	27.16	26.58	23.50	30.96	27.04	19.69	19.19
FeO	0.09	0.36	0.30	0.24	0.62	0.55	0.62	0.55	0.62	0.41	0.40	0.37	0.41	0.22
MnO	0.03	0.03	0.00	0.01	0.03	0.00	0.04	0.00	0.03	0.03	0.01	0.00	0.06	0.02
MgO	0.07	0.01	0.02	0.01	0.05	0.07	0.03	0.07	0.05	0.02	0.04	0.03	0.29	0.28
CaO	9.51	4.40	4.96	11.95	9.33	10.30	4.87	10.3	9.33	5.71	14.44	9.90	0.57	0.35
Na ₂ O	5.51	8.81	8.54	4.55	5.52	5.05	8.08	5.05	5.52	7.93	3.28	5.66	11.07	11.13
K ₂ O	0.20	0.37	0.49	0.13	0.82	0.70	0.44	0.70	0.82	0.77	0.20	0.37	0.11	0.05
Total	96.84	99.19	99.00	98.17	99.54	99.34	49.15	99.34	99.54	100.07	99.31	99.31	99.16	97.48
<i>Number of ions on the basis of 32 oxygen</i>														
Si	10.10	11.09	11.10	9.69	10.24	10.09	4.65	10.09	10.24	10.98	9.20	10.14	11.82	11.89
Al	5.88	4.88	4.83	6.24	5.67	5.82	11.32	5.82	5.67	4.93	6.71	5.78	4.11	4.06
Ti	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.02	0.00
Fe ²⁺	0.01	0.05	0.05	0.04	0.09	0.08	0.21	0.08	0.09	0.06	0.06	0.06	0.06	0.03
Mn	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00
Mg	0.02	0.00	0.01	0.00	0.01	0.02	0.02	0.02	0.01	0.01	0.01	0.01	0.08	0.08
Ca	1.89	0.84	0.95	2.36	1.81	2.01	2.13	2.01	1.81	1.09	2.85	1.93	0.11	0.07
Na	1.98	3.06	2.97	1.63	1.94	1.78	6.38	1.78	1.94	2.74	1.17	1.99	3.80	3.88
K	0.05	0.08	0.11	0.03	0.19	0.16	0.23	0.16	0.19	0.18	0.05	0.09	0.02	0.01
Total	19.95	20.00	20.02	20.00	19.97	19.96	24.96	19.97	19.96	19.99	20.05	19.99	20.03	20.02
%Ab	50.60	76.70	73.60	40.50	49.20	45.10	73.00	45.10	49.20	68.40	28.80	49.80	96.62	98.00
%An	48.20	21.2	23.60	58.80	46.00	50.80	24.30	50.80	46.00	27.20	70.10	48.10	2.77	1.70
%Or	1.20	2.10	2.80	0.80	4.80	4.10	2.60	4.10	4.80	4.40	1.20	2.10	0.61	0.30

and minor titanite. Plagioclase occurs as large, euhedral to subhedral crystals (Fig. 4b) which range in composition from An₂₇ to An₇₀, with Or content rarely exceeding 4 mol% (Table 2). Clinopyroxene is subhedral and was altered to chlorite and amphibole in some samples. Clinopyroxenes have diopsidic composition and are characterized by 0.32–0.51 wt.% TiO₂, 0.94–1.95 wt.% Al₂O₃ and $X_{Mg} [= Mg / (Mg + Fe_{tot})] = 0.71$ (Table 3).

The pyroxene-quartz-monzodiorites show a granophyric texture that contains quartz and K-feldspar in characteristic angular intergrowths. Porphyritic and granophyric textures are common in the ZZM (Fig. 4a) and indicate that these intrusions were emplaced at a

shallow level as hot, dry magmas. Opaque minerals are rare and have a euhedral shape. In some parts of this pluton, the composition of rocks tends to be quartz-monzonite.

4.2. Zaker pluton

The Zaker pluton shows mainly pyroxene-quartz-monzonite compositions. The coarse- to medium-grained rocks, pinkish-gray in color, are porphyritic and granular rocks. They contain 1–4 cm large phenocrysts of idiomorphic plagioclase (An₂₁ to An₅₈; Table 2) and subhedral K-feldspar in a medium- to fine-grained groundmass

Table 3
Microprobe analyses (wt.%) of clinopyroxenes in the representative rock types of the Tarom plutons.

Sample	Kuh Tabar micro quartz monzodiorite			Zanjan quartz monzodiorite			Zaker quartz monzonite		
SiO ₂	52.31	51.65	53.76	53.11	52.77	53.48	54.06	53.86	53.67
TiO ₂	0.52	0.44	0.42	0.32	0.51	0.44	0.10	0.11	0.11
Al ₂ O ₃	1.66	2.67	1.31	0.94	1.95	1.55	0.35	0.41	0.32
FeO	10.93	10.10	11.88	9.39	9.85	10.03	7.60	7.23	7.97
Cr ₂ O ₃	0.00	0.00	0.00	0.02	0.01	0.00	0.00	0.02	0.02
MnO	0.51	0.37	0.53	0.72	0.37	0.42	0.46	0.47	0.43
MgO	13.38	14.03	13.28	13.85	13.96	14.11	14.88	15.10	14.83
CaO	19.64	18.97	19.52	22.91	20.32	19.76	22.41	22.23	21.61
Na ₂ O	0.37	0.28	0.32	0.35	0.27	0.32	0.31	0.32	0.38
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.32	98.51	101.02	101.61	100.01	100.11	100.17	99.76	99.34
<i>Number of ions on the basis of 6 oxygen</i>									
Si	1.98	1.96	2.00	1.95	1.97	2.00	2.00	2.00	2.00
Al ^{IV}	0.03	0.05	0.00	0.04	0.03	0.01	0.00	0.00	0.00
Al ^{VI}	0.05	0.07	0.06	0.00	0.06	0.06	0.02	0.02	0.01
Ti	0.02	0.01	0.02	0.02	0.01	0.01	0.00	0.00	0.00
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.35	0.32	0.37	0.29	0.31	0.31	0.24	0.23	0.25
Mn	0.02	0.01	0.02	0.02	0.01	0.01	0.01	0.02	0.01
Na	0.03	0.02	0.02	0.03	0.02	0.02	0.02	0.02	0.03
Ca	0.80	0.77	0.78	0.90	0.81	0.81	0.89	0.88	0.86
Mg	0.75	0.79	0.74	0.76	0.78	0.79	0.82	0.84	0.83
Total	4.03	4.00	4.01	4.01	4.00	4.02	4.00	4.01	4.00
X _{Mg}	0.68	0.71	0.67	0.72	0.71	0.71	0.77	0.79	0.77
%Wo	41.02	40.65	40.93	45.75	42.12	41.02	44.85	44.61	44.28
%En	38.89	41.83	38.75	38.48	40.27	40.76	41.44	42.15	42.28
%Fs	18.70	17.52	20.32	15.77	16.59	17.02	12.59	12.07	13.44

Table 4

Major (wt.%) and trace (ppm) element concentrations of the Tarom magmatic rocks (plutonic and volcanic).

Rock type	Sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Ba	Rb	Sr	Y	Zr	Nb	Th	Pb	Ga	Zn	Cu
Z.Qmon	M 25	61.00	0.94	15.02	7.09	0.17	2.42	5.60	2.96	4.47	0.31	968	98.0	320	34.1	217	13.8	10.1	9.0	16.7	57.0	33.0
Z.Qmon	M 39	63.73	1.11	16.03	3.60	0.13	2.53	7.06	4.93	0.82	0.05	279	23.9	377	35.3	256	15.7	8.3	6.0	17.1	36.0	15.0
Z.Qmon	M 47	62.41	1.06	15.40	6.10	0.13	1.86	5.27	3.30	4.25	0.20	996	95.0	341	33.8	254	16.2	13.4	15.0	17.2	62.0	16.0
Z.Qmon	N-A1	61.95	1.10	15.49	3.55	0.18	3.58	7.11	4.53	1.37	0.02	327	56.4	462	35.6	169	7.9	4.5	4.4	15.6	19.9	4.3
Z.Qmon	N-SD19	60.63	0.94	16.00	4.76	0.10	2.29	6.31	3.72	3.44	0.37	854	66.4	395	38.5	247	13.3	9.7	5.5	16.4	32.9	19.0
Z.Qmon	GN1	62.47	0.71	15.63	6.35	0.11	1.88	5.50	2.72	3.94	0.21	923	99.5	315	29.4	236	13.5	9.6	6.0	16.2	55.0	57.0
Z.Qmon	G7	61.89	0.78	15.98	5.17	0.09	2.27	6.55	3.38	2.72	0.66	1150	94.9	346	39.5	192	14.7	10.5	7.0	15.6	72.0	31.0
Z.Qmon	SD86	66.73	0.81	15.34	6.11	0.13	2.11	2.59	2.50	2.69	0.31	1145	110	340	36.0	246	16.1	13.5	8.0	15.5	54.0	16.0
Z.Qmon	SD92	65.28	0.76	15.00	6.31	0.10	1.27	2.51	4.09	3.92	0.46	1060	139	238	45.0	355	20.2	16.6	20.0	16.3	102.0	39.0
Z.Qmon	G20	66.26	0.85	15.13	6.11	0.15	2.47	2.73	2.55	2.63	0.42	1280	117	268	37.4	256	14.7	12.0	25.0	15.6	67.0	17.0
M.M.S	M 60	68.30	0.56	14.96	1.97	0.06	2.02	3.77	6.08	2.22	0.05	448	50.9	318	17.7	130	9.7	9.4	8.0	17.2	33.0	6.0
M.M.S	N-SD4	69.36	0.69	14.25	2.39	0.06	0.81	2.41	4.96	3.36	0.03	928	62.0	329	19.2	235	9.2	7.9	6.2	14.2	35.4	21.2
M.M.S	N-M65	69.68	0.59	14.53	3.60	0.06	0.44	2.25	3.97	3.98	0.12	1029	72.7	365	28.6	248	19.9	19.3	6.8	14.8	20.5	6.4
M.M.S	N-M16	70.88	0.53	14.16	3.22	0.05	0.37	1.31	3.03	5.92	0.12	1211	139	266	26.5	247	18.4	16.4	11.3	13.5	27.6	5.6
En-M.M.S	N-M16 EN	70.18	0.65	13.35	3.73	0.06	0.58	1.58	2.85	5.95	0.15	785	212	125	36.0	423	19.4	31.5	14.8	13.8	52.2	6.4
Z.Q.M	N-T19	61.05	0.56	18.35	4.76	0.10	1.54	5.50	3.53	3.76	0.24	723	109	496	25.4	175	11.6	9.7	14.8	16.7	87.0	73.6
Z.Q.M	N-T38	59.72	0.64	16.86	5.95	0.09	2.73	5.67	3.25	3.49	0.31	806	103	436	28.5	176	12.2	10.1	8.2	16.4	37.3	182.0
Z.Q.M	N-T52	57.12	0.80	16.49	7.84	0.13	3.05	6.26	3.18	3.13	0.32	612	88.6	448	26.8	186	11.7	8.2	9.2	16.5	70.0	88.6
Z.Q.M	N-T57	57.81	1.10	15.04	8.87	0.17	3.65	5.88	2.81	2.47	0.33	597	56.9	299	33.7	200	13.4	8.2	7.5	16.3	85.6	82.5
Z.Q.M	N-Z8	57.78	1.21	14.02	7.72	0.13	3.94	9.18	3.13	2.00	0.41	482	38.8	336	61.8	95.9	16.8	12.3	5.0	15.8	32.1	12.6
En-Z.Q.M	N-T52 En	50.33	0.94	17.31	12.66	0.16	4.74	7.16	3.48	0.65	0.18	222	22.9	459	16.0	68.8	5.6	3.5	39.6	18.7	215.0	950.0
K.T.QMD	N-D26	63.58	0.74	16.36	5.45	0.06	0.94	3.07	2.77	5.29	0.23	1221	137	477	31.0	259	20.2	11.5	13.1	16.1	42.2	17.3
K.T.QMD	N-D7	61.69	0.81	16.13	5.68	0.18	1.88	4.81	2.76	4.36	0.27	855	123	383	32.2	253	19.5	11.3	18.0	17.4	144.0	63.5
K.T.QMD	N-D9	62.31	0.84	15.96	5.75	0.12	1.96	4.49	2.86	4.35	0.26	810	127	328	32.2	256	20.4	11.4	29.9	17.0	83.4	44.0
K.T.QMD	N-GH2	60.98	0.71	15.50	6.60	0.11	2.66	5.01	2.98	3.75	0.21	787	111	375	30.1	192	13.2	11.5	11.6	16.0	79.8	94.0
Volc	N-T1	64.35	0.49	17.53	3.57	0.08	1.01	4.43	3.55	3.98	0.13	779	116	385	26.5	224	13.0	13.3	14.0	16.0	45.8	31.1
Volc	N-T48	64.67	0.51	17.56	3.65	0.08	1.05	4.39	3.59	3.94	0.17	813	117	378	27.2	219	13.1	12.4	18.4	16.0	51.9	19.4
Volc	M 120	49.41	0.91	19.54	9.33	0.10	4.81	11.63	3.27	0.75	0.24	278	18.1	544	16.5	55.0	3.2	1.7	5.0	19.2	30.0	11.0
Volc	M 126	57.72	0.96	15.65	8.35	0.10	3.85	5.65	3.04	4.32	0.36	865	109	326	30.4	190	11.9	8.7	5.0	16.7	28.0	35.0
Volc	M 130	58.32	1.15	16.77	9.59	0.14	3.00	4.03	3.95	2.64	0.39	689	64.8	509	33.9	165	10.1	7.0	7.0	19.2	121.0	5.0
Volc	N-T24	62.53	0.45	19.53	3.30	0.07	0.95	5.67	3.69	3.23	0.14	745	72.8	520	22.4	178	10.7	10.4	13.5	16.9	40.5	23.8
Volc	N-T41	64.27	0.52	17.72	3.79	0.07	1.08	4.54	3.63	3.84	0.17	780	107	399	27.0	207	12.5	11.9	12.2	16.1	60.8	48.3

Rock type	Sample	Ni	V	Cr	Hf	Cs	Ta	Co	U	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	La/Yb	K ₂ O/Na ₂ O
Z.Qmon	M 25	6.0	119	130	6.0	2.1	0.8	7.7	2.6	24.4	49.7	6.5	27.2	6.4	1.5	6.7	1.1	6.2	1.3	3.8	0.6	3.7	0.6	6.59	1.51
Z.Qmon	M 39	5.0	55.0	100	7.0	1.9	1.1	6.8	1.6	12.8	28.7	4.3	19.2	5.5	1.5	5.9	1.1	6.7	1.5	4.2	0.6	4.4	0.7	2.91	0.17
Z.Qmon	M 47	5.0	103	80	7.1	0.8	1.1	8.5	2.4	19.1	40.6	5.6	23.9	5.8	1.5	6.4	1.1	6.4	1.4	4.0	0.6	4.0	0.6	4.78	1.29
Z.Qmon	N-A1	3.3	38.8	25.0	4.4	3.5	0.8	6.7	0.6	12.3	25.8	3.5	16.8	4.9	1.2	5.7	0.9	6.0	1.3	3.7	0.6	3.9	0.6	3.15	0.30
Z.Qmon	N-SD19	5.1	121	21.4	6.2	1.1	0.8	8.3	1.6	26.5	56.9	7.0	30.6	7.1	1.6	7.2	1.1	6.6	1.4	3.9	0.6	3.9	0.6	6.79	0.92
Z.Qmon	GN1	<5	72.0	30	6.4	2.2	0.9	8.3	2.4	18.4	38.0	4.7	20.6	5.1	1.3	5.3	0.8	5.3	1.1	3.2	0.5	3.5	0.5	5.26	1.45
Z.Qmon	G7	8.0	108	50	5.4	2.1	0.9	9.9	1.8	30.7	68.4	8.7	36.4	8.2	1.6	8.5	1.2	7.2	1.4	4.2	0.6	4.0	0.6	7.68	0.80
Z.Qmon	SD86	<5	104	40	6.7	2.8	1.0	8.6	2.0	27.4	62.5	8.0	34.0	7.6	1.7	7.6	1.1	6.6	1.3	3.9	0.6	3.7	0.6	7.41	1.08
Z.Qmon	SD92	<5	53.0	80	9.6	2.0	1.3	9.7	3.5	40.4	82.6	9.8	40.4	8.8	1.7	9.0	1.4	8.4	1.7	5.0	0.7	4.9	0.7	8.24	0.96
Z.Qmon	G20	<5	75.0	40	6.8	1.3	0.9	5.8	2.7	22.4	45.3	5.7	24.4	6.0	1.3	6.7	1.1	6.8	1.4	4.2	0.6	4.1	0.6	5.46	1.03
M.M.S	M 60	5.0	42.0	70	3.7	0.5	0.6	3.3	1.3	10.6	20.4	2.7	11.2	2.8	0.9	3.0	0.5	3.2	0.7	2.0	0.3	2.1	0.4	5.05	0.37
M.M.S	N-SD4	1.5	30.5	4.4	5.9	1.3	1.0	3.8	1.3	10.7	18.2	2.1	9.3	2.6	1.2	2.9	0.5	3.1	0.7	2.0	0.3	2.3	0.4	4.65	0.68
M.M.S	N-M65	2.5	44.9	2.2	6.9	0.7	1.5	3.6	1.9	17.6	30.6	3.5	15.5	4.0	1.2	4.4	0.7	4.5	0.9	2.8	0.4	2.9	0.4	6.07	1.00
M.M.S	N-M16	2.0	28.2	1.7	6.8	1.5	1.2	3.3	2.3	20.4	37.0	4.3	17.6	4.1	1.2	4.4	0.7	4.3	0.9	2.7	0.4	2.8	0.4	7.29	1.95
En-M.M.S	N-M16 EN	1.4	50.3	1.6	10.8	2.9	1.4	5.2	7.5	40.8	79.7	8.7	33.0	6.6	1.0	6.5	1.0	5.9	1.2	3.5	0.6	3.7	0.6	11.03	2.09
Z.Q.M	N-T19	3.5	83.9	6.6	4.4	2.3	0.8	9.6	2.6	26.6	51.0	5.8	23.1	4.9	1.3	4.7	0.7	4.2	0.9	2.5	0.4	2.6	0.4	10.23	1.07
Z.Q.M	N-T38	10.4	125	20.5	4.5	1.9	0.8	16.8	2.4	27.4	53.2	6.1	25.2	5.5	1.3	5.3	0.8	4.7	1.0	2.8	0.4	2.8	0.4	9.79	1.07
Z.Q.M	N-T52	9.6	175	22.9	4.6	1.5	0.7	19.3	2.1	23.4	46.2	5.4	22.7	5.0	1.3	5.0	0.7	4.5	0.9	2.7	0.4	2.7	0.4	8.67	0.98
Z.Q.M	N-T57	15.2	193	27.2	5.0	1.2	0.8	14.3	2.3	25.6	52.3	6.2	26.5	6.0	1.4	6.1	0.9	5.6	1.2	3.4	0.5	3.4	0.5	7.53	0.88
Z.Q.M	N-Z8	6.8	198	35.4	2.7	0.4	1.4	12.8	1.4	32.4	75.4	9.9	43.0	10.6	1.7	11.1	1.7	10.6	2.2	6.2	0.9	6.2	0.9	5.23	0.64
En-Z.Q.M	N-T52 En	37.6	261	17.9	1.8	0.7	0.4	41.8	1.6	16.3	30.0	3.3	13.4	2.8	1.3	2.8	0.4	2.5	0.6	1.6	0.3	1.9	0.3	8.58	0.19
K.T.QMD	N-D26	3.3	100	7.2	6.2	4.7	1.3	1.2	3.3	31.5	59.9	6.7	27.1	5.7	1.5	5.6	0.8	5.1	1.1	3.1	0.5	3.1	0.5	10.16	1.91
K.T.QMD	N-D7	5.5	98.9	11.2	6.2	2.2	1.3	11.3	3.2	34.2	66.7	7.5	29.8	6.2	1.5	6.1	0.9	5.5	1.1	3.2	0.5	3.2	0.5	10.69	1.58
K.T.QMD	N-D9	5.2	106	11.6	6.3	2.1	1.3	11.3	3.2	35.2	68.3	7.7	30.9	6.5	1.5	6.2	0.9	5.5	1.1	3.2	0.5	3.3	0.5	10.67	1.52
K.T.QMD	N-GH2	8.7	143	19.8	5.0	1.5	0.9	16.1	3.2	27.0	53.1	6.1	24.8	5.4	1.3	5.4	0.8	5.1	1.1	3.0	0.5	3.1	0.5	8.71	1.26
Volc	N-T1	2.2	52.8	3.1	5.7	2.3	0.9	6.0	4.2	25.7	50.2	5.6	22.4	4.7	1.2	4.7	0.7	4.2	0.9	2.6	0.4	2.7	0.4	9.52	1.12
Volc	N-T48	2.9	53.8	4.6	5.5	2.5	0.9	5.9	2.6	26.0	50.6	5.7	23.1	4.9	1.2	4.8	0.7	4.4	0.9	2.7	0.4	2.8	0.4	9.29	1.10
Volc	M 120	20.0	334	90	1.7	0.3	0.2	19.8	0.5	10.4	21.7	3.0	12.9	3.1	1.0	3.5	0.6	3.3	0.7	1.9	0.3	1.8	0.3	5.78	0.23
Volc	M 126	18.0	184	100	5.4	0.7	0.8	16.3	5.4	21.5	45.1	5.9	24.5	5.5	1.4	6.0	1.0	5.8	1.2	3.5	0.6	3.7	0.6	5.81	1.42
Volc	M 130	16.0	217	70	4.5	0.8	0.6	20.1	2.0	31.7	58.4	7.7	31.2	6.8	1.8	7.2	1.1	6.3	1.3	3.8	0.5	3.5	0.5	9.06	0.67
Volc	N-T24	2.0	46.5	2.6	4.5	1.3	0.7	5.3	3.1	23.1	43.4	4.9	19.5	4.1	1.3	4.0	0.6	3.6	0.7	2.2	0.3	2.3	0.4	10.04	0.88
Volc	N-T41	2.5	56.5	5.1	5.3	2.3	0.8	6.6	3.1	25.4	49.7	5.7	22.6	4.9	1.3	4.8	0.7	4.4	0.9	2.7	0.4	2.7	0.4	9.41	1.06

Abbreviations are: Z.Qmon: Zaker Quartz monzonite; K.T.QMD: Kuh-Tabar quartz-monzodiorite; M.M.S: Morvarid Monzogranite to Syenogranite; En: Enclave; Z.Q.M: Zanjan quartz-monzodiorite; Volc: Eocene volcanic rocks.

which is composed of K-feldspar (36–40%), plagioclase (38–42%), quartz (10–12%) and diopsidic clinopyroxene (7–9%) (Table 1). Plagioclase and K-feldspar typically occur in approximately equal proportions. Clinopyroxene show diopside composition and $X_{Mg} = 0.77$ (Table 3). Quartz occurs in the anhedral form, and a granophyric texture is well developed between the large K-feldspar crystals (intergrowth of quartz and K-feldspar).

Apatite, magnetite and zircon can be found as accessory phases, while epidote, chlorite, actinolite, sericite, clay minerals and carbonate are secondary minerals. Most of the plagioclases are altered to epidote, clay minerals, sericite and calcite. Some of the K-feldspars show a perthitic texture. Clinopyroxene occurs between the plagioclases and K-feldspars. In parts, small euhedral clinopyroxene crystals can be observed within plagioclase grains (Fig. 4b). Clinopyroxene is partially replaced by actinolite and chlorite during subsolidus reactions.

Electronic microprobe and reflective light microscope studies show that idiomorphic crystals of magnetite are also abundant in these rocks. The presence of primary magnetite in the pyroxene and plagioclase indicates high oxygen fugacity in the parent magma (Sack et al., 1980). In the study area, there are some iron oxide-apatite veins, which are mostly hosted within the Zaker quartz-monzonite pluton. Field and structural studies have shown that these plutonic bodies are restricted between two major strike-slip faults with a NW–SE orientation. Between these strike-slip faults, there are some parallel ENE–WSW trending faults (Fig. 2). These secondary faults may be strike-slip faults in their origin, which host the magnetite-apatite ore bodies.

4.3. Morvarid pluton

Monzogranites to syenogranites occur on both sides of the Zaker pluton as small stocks (Fig. 2). These represent a fine- to medium-grained leucocratic rock, with porphyritic and granophyric textures. Monzogranite samples contain quartz (25–30%), K-feldspar (43–47%) and plagioclase (25–27%) with biotite (4–6%) (Table 1). Plagioclase is mostly sodic and ranges in composition from An_2 to An_5 , with Or content rarely exceeding 0.6 mol% (Table 2). Accessory phases are magnetite, apatite and zircon.

4.4. Kuh-Tabar pluton

This intrusive body consists of fine-grained porphyritic rocks with a quartz-monzodiorite composition. The Kuh-Tabar pluton consists of up to 2 mm-long zoned and twinned plagioclase grains (An_{24} to An_{51} ; Table 2), clinopyroxene and some minor quartz and K-feldspar. The clinopyroxene grains form subhedral prisms, are up to 1 mm in length, and most of them are altered to epidote, chlorite and calcite.

Clinopyroxenes are generally diopside to augite and characterized by $X_{Mg} = 0.67$ –0.71 (Table 3). The cores of some of the larger plagioclase grains contain small inclusions of clinopyroxene. Subordinate phases are magnetite, apatite, and secondary sericite, epidote, chlorite and calcite. There are some xenocrysts in the Kuh-Tabar intrusive body, which seem to be related to the Zaker plutonic rocks. The xenocrysts have a mineralogical composition similar to the Zaker samples, including plagioclase, K-feldspar, clinopyroxene and magnetite (Fig. 4c). The xenocrysts contain minerals of smaller size than the phenocrysts in the Kuh-Tabar plutonic rocks and show resorbed margins (Fig. 4c).

4.5. Volcanic rocks

The host rocks of all the plutons are Eocene volcanics. The dark gray and fine-grained porphyritic andesites are predominantly composed of plagioclase, clinopyroxene and quartz with accessory magnetite and minor apatite. Phenocrysts are mainly plagioclase and clinopyroxene. Plagioclase appears as anhedral to subhedral plates, are zoned and are

often altered to sericite, epidote and calcite. Clinopyroxene crystals show a euhedral to subhedral shape and are often altered to biotite, amphibole, chlorite, calcite and opaque minerals. Quartz occurs as both anhedral to subhedral crystals and as a late interstitial phase. Minor alteration products are sericite, chlorite, epidote, and calcite.

5. Analytical methods

5.1. Microprobe analyses

Mineral compositions were determined at the University of Salzburg, Austria using a JEOL 8600 superprobe. A 15 kV electron beam accelerating voltage, and 40 nA beam current per analysis were applied. Electron beam spot diameter used was 5 μ m diameter (for silicate and apatite). Synthetic and natural standards were used (SiO_2 , Al_2O_3 , NaCl, KCl, apatite (P, Ca, F for apatite analyses), MgO, Cr_2O_3 , TiO_2 , FeO, MnO, wollastonite (Ca) counting time on peak 10 s (for F 15 s) and on background 5 s). System ZAF 4 software was used to convert X-ray spectra obtained from the specimen into chemical analyses. Chemical compositions and structural formula of minerals are listed in Tables 2 and 3.

5.2. Whole-rock major and trace element analyses

Fresh chips of whole rock samples were powdered to 200 meshes using a tungsten carbide ball mill, and preserved in a desiccator for analyses after being dried in an oven at 105 °C for 2 h. Major and trace element compositions were analyzed by means of XRF (Rigaku RIX 2100) and ICP-MS (PE 6100 DRC), respectively, at the State Laboratory of Continental Dynamics, Northwest University, Xi'an, China. For major element analysis, 0.5 g sample powders were mixed with 3.6 g $Li_2B_4O_7$, 0.4 g LiF, 0.3 g NH_4NO_3 and minor LiBr in a platinum pot, and melted to a glass disc in a high frequency melting instrument prior to analysis. For trace element analysis, sample powders were digested using a HF + HNO_3 mixture in high-pressure Teflon bombs at 190 °C for 48 h. Analyses of USGS and Chinese national rock standards (BCR-2, GSR-1 and GSR-3) indicate that both analytical precision and accuracy for major elements are generally better than 5% for most of the trace elements except for transition metals for which they are better than 2% and 10%, respectively. We also include the geochemical analyses presented by Nabatian (2008) from the ZZM intrusive rocks, from which major and trace element analyses by Inductively Coupled Plasma-Mass Spectrometer (ICP-MS) were performed at the ALS-Chemex Laboratory in Canada.

5.3. Sr, Nd and Pb isotopic analysis

For Rb–Sr and Sm–Nd isotope geochemistry details of analytical techniques, accuracy and precision can be found in von Quadt et al. (1999). For the isotope analyses, ~100 mg of whole-rock powder was dissolved in HF and HNO_3 , followed by Pb and Sr separation by exchange chromatography techniques on PP columns with Sr spec Eichrom resin. Strontium and Pb isotopes were analyzed on a Finnigan MAT 262 by static mode TIMS at ETH Zurich. Strontium was loaded with a Ta emitter on Re filaments. The measured $^{87}Sr/^{86}Sr$ ratios were normalized to an $^{88}Sr/^{86}Sr$ value of 8.37521. The mean $^{87}Sr/^{86}Sr$ value of the NBS 987 standard obtained during the period of measurements was 0.710252 ± 12 (2σ , $n = 18$). Lead was loaded with a silica gel on Re filaments. A mass fractionation correction of 1.1‰ per atomic mass unit was applied, based on replicate analyses of the NBS 982 reference material. Age-corrected Sr and Pb isotope ratios were calculated using Rb, Sr, U, Th and Pb concentrations determined by Laser Ablation (LA)-ICP-MS.

Six samples from the Tarom plutonic complex have also been analyzed in the Nanjing University, China. Isotope geochemistry details of analytical techniques can be found in Zhao et al. (2012).

5.4. U–Pb dating

The U–Pb analytical techniques largely follow those described in Liu et al. (2008). Zircons were dated in-situ on an excimer (193 nm wave length) LA-ICP-MS at the State Key Laboratory of Continental Dynamics, Northwest University, Xi'an, China. The ICP-MS used is an Agilent 7500a (with shield torch). The unique shield torch increases analytical sensitivity by a factor of >10 (for example, 4500 cps/ppm ^{238}U at a spot size of $40\text{ }\mu\text{m}$ and laser frequency of 10 Hz), which is important for LA-ICP-MS. The GeoLas 200 M laser ablation system (MicroLas, Göttingen, Germany) was used for the laser ablation experiments. Helium was used as carrier gas. The used spot size and laser frequency

were $40\text{ }\mu\text{m}$ and 10 Hz, respectively. The data acquisition mode was peak jumping (20 ms per isotope each cycle). Raw count rates were measured for ^{29}Si , ^{204}Pb , ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th and ^{238}U . Uranium, Th and Pb concentrations were calibrated using ^{29}Si as an internal standard and NIST SRM 610 as the reference standard. Each analysis consists of 30 s gas blank and 40 s signal acquisition. High-purity argon was used together with a custom helium filtration column, which resulted in ^{204}Pb and ^{202}Hg being less than 100 cps in the gas blank. Therefore, the contribution of ^{204}Hg to ^{204}Pb was negligible and no correction was made. $^{207}\text{Pb}/^{206}\text{Pb}$, $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{208}\text{Pb}/^{232}\text{Th}$ ratios, calculated using GLITTER 4.0 (Macquarie University), were corrected for both instrumental mass bias and depth-dependent elemental and

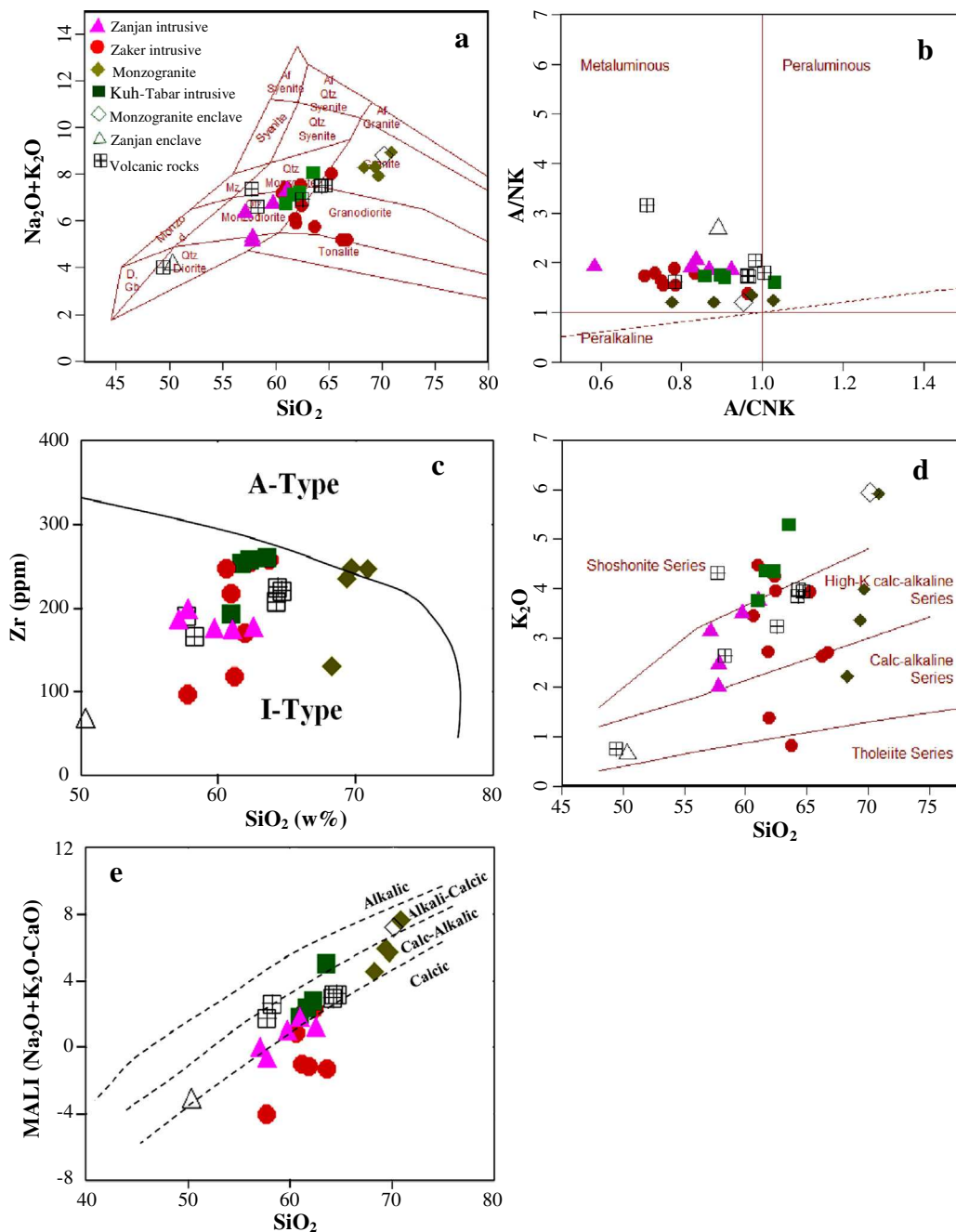


Fig. 5. Geochemical variation diagrams for the Tarom complex emphasizing some chemical features that distinguish the granitoid rocks. (a) Rock classification diagram (Middlemost, 1994) for the Tarom complex. (b) A/NK (molar) versus A/CNK (after Maniar and Piccoli, 1989) diagram for the pluton. (c) Zr vs. SiO_2 diagram proposed by Collins et al. (1982). (d) MALI- SiO_2 diagram (Frost et al., 2001). (e) K_2O versus SiO_2 diagram for the samples with lines separating the tholeiitic, calc-alkaline, high-K calc-alkaline and shoshonitic series of Peccerillo and Taylor (1976).

isotopic fractionation using Harvard zircon 91500 as an external standard. The ages were calculated using ISOPLOT 3 (Ludwig, 2003). Our measurement of TEMORA 1 as an unknown yielded a weighted $^{206}\text{Pb}/^{238}\text{U}$ age of 415 ± 4 Ma (MSWD = 0.112, $n = 24$) (Yuan et al., 2004), which is in good agreement with the recommended ID-TIMS age of 416.75 ± 0.24 Ma (Black et al., 2003). Analytical details for age and trace and rare earth element determinations of zircons are reported in Yuan et al. (2004). Common Pb corrections were made following the method of Andersen (2002). Because measured ^{204}Pb usually accounts for <0.3% of the total Pb, the correction is insignificant in most cases.

6. Geochemical characteristics

6.1. Major and trace elements

Whole-rock analyses of major and trace elements from all samples are reported in Table 4. We use the classification diagram of Middlemost (1994) for characterizing the rock types. The samples plot in the fields of quartz-diorite, quartz-monzodiorite, quartz-monzonite and granite (Fig. 5a). Mafic enclaves enclosed in the quartz-monzonite and monzogranite are quartz-diorite and monzogranite in

composition, respectively. The results of geochemical classification of the Tarom samples are in good agreement with petrographic signatures.

All of the plutonic and volcanic rocks exhibit metaluminous (Fig. 5b) and I-type (Fig. 5c) geochemical characters with ASI [= molar $\text{Al}_2\text{O}_3 / (\text{CaO} + \text{K}_2\text{O} + \text{Na}_2\text{O})$] ranging from 0.6 to 1.05. Most of the plutonic and volcanic samples show a high potassic to shoshonitic tendency on the K_2O – Na_2O diagram (Fig. 5d). Based on Irvine and Baragar (1971) diagrams (not shown), all of the samples exhibit calc-alkaline affinity. According to the Peacock's calc-alkalic index represented in the MALI– SiO_2 diagram (Frost et al., 2001; MALI-modified lime-alkali index = $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$ wt.%; Fig. 5e) most of the ZMZ samples and their related enclaves plot within or near the calcic to calc-alkali fields, following the trend of the field boundaries. The Kuh-Tabar, volcanic and some of the Morvarid monzogranite samples lie across the field boundaries evolving from calc-alkali to alkali-calcic series.

Major element variations are illustrated in major oxide multi-element variation diagrams (Fig. 6). The Tarom magmatic rocks exhibit a range in SiO_2 content from 57.1 to 70.8 wt.%. An enclave in the Zanjan pluton has also been analyzed showing 50 wt.% SiO_2 content. The multi-element variation diagrams (Fig. 6) show decreasing abundances of

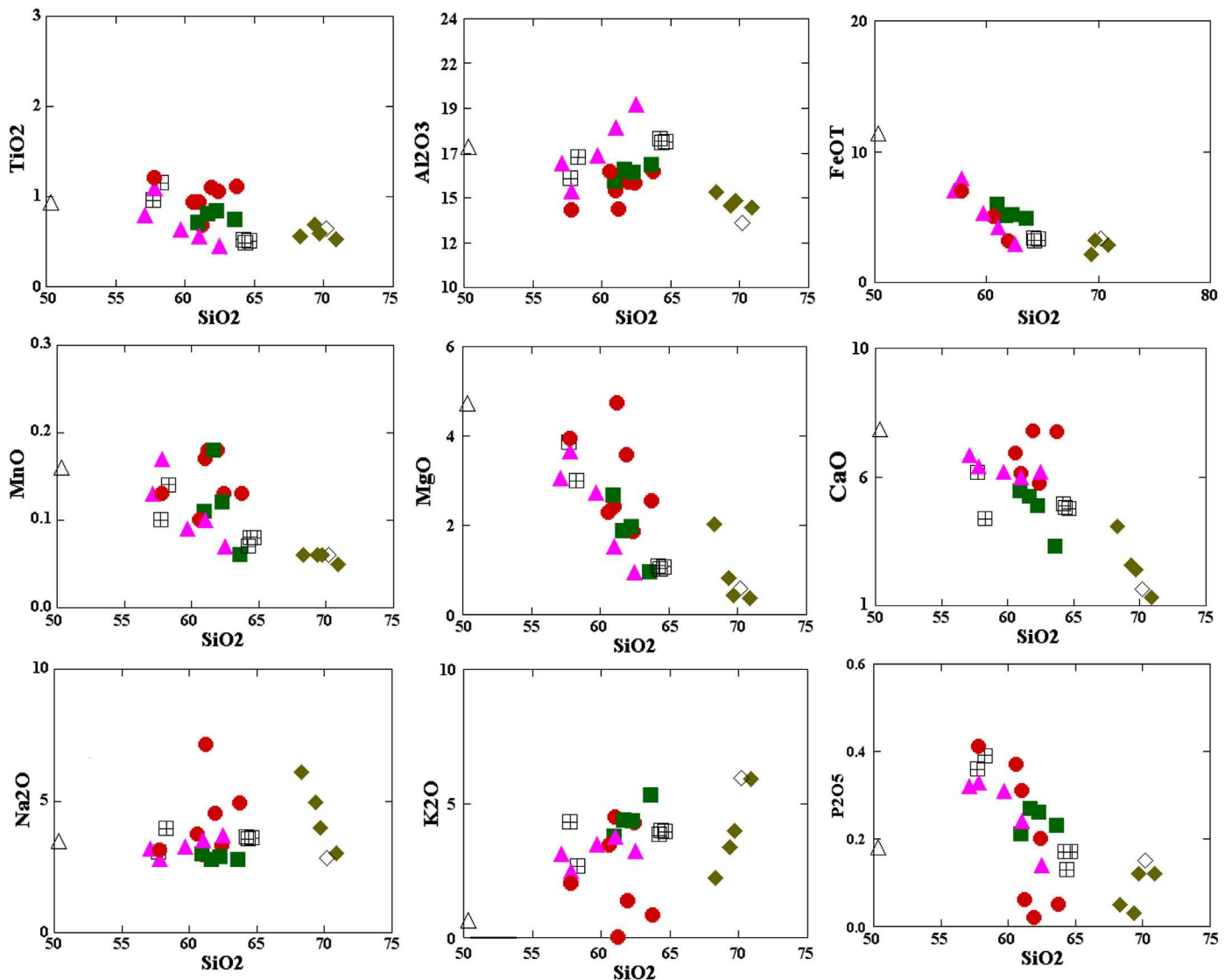


Fig. 6. Selected major oxides vs. SiO_2 (wt.%) contents for the Tarom complex.

FeO, MgO, MnO, CaO, P₂O₅, Al₂O₃ and TiO₂ contents with increasing SiO₂, whereas K₂O and Na₂O increase.

Tarom plutonic samples do not show a meaningful correlation from mafic to felsic samples for most elements (Fig. 6). The rocks are potassic as shown in the K₂O vs. SiO₂ diagram (Fig. 6) and have higher average values of K₂O than those of Urumieh-Dokhtar and Andean arc plutonic rocks. Samples from the Kuh-Tabar quartz-monzonite have higher K₂O contents and lie within the shoshonitic field (Fig. 5d). In the K₂O vs. SiO₂ diagram, the monzogranite samples show a direct trend with a relatively constant SiO₂. This trend can also be seen in the other mobile elements such as Na₂O, Rb, Ba, Th and La, which may show potassic alteration in these samples. Kuh-Tabar samples show linear trends with a negative slope in the TiO₂, MnO, MgO, CaO, Fe₂O₃ and P₂O₅ contents and a narrower positive trend in Al₂O₃ and Na₂O diagrams. However, K₂O evolution is more complex. The Tarom rocks have the characteristics of high alkalis (K₂O + Na₂O > 5.0%), high K₂O/Na₂O ratio > 0.4, low TiO₂ contents (mostly < 0.9%), and high and variable Al₂O₃ contents (mostly 13–18%), therefore ruling out the ultrapotassic series.

Variation diagrams of SiO₂ versus selected trace elements (Fig. 7) display a general decreasing of most compatible elements and increasing large ion lithophile elements (LILEs) with increasing SiO₂ contents. However, some samples show complex variations such as Sr and Rb which may be related to some alteration of these samples. Variation of

V and Ni contents in each pluton of the Tarom complex is consistent with clinopyroxene and magnetite fractionation in these plutons suggesting separate fractional crystallization history for Zanjan, Zaker and Kuh-Tabar plutons. In most variation diagrams (Figs. 6 and 7), Tarom plutons show separate evolution and do not follow a common trend from mafic to felsic samples.

All the Tarom samples display similar trace element abundance patterns, with enrichment in LILE (e.g., Cs, Rb, Th and K) and pronounced negative anomalies in high field strength elements (HFSEs) such as Ti and Nb in the chondrite normalized spider diagrams (Thompson, 1982; Fig. 8). Rare earth element (REE) patterns for the complex are represented in Fig. 9. All the samples are significantly enriched in light REE (LREE), and strongly depleted in heavy REE (HREE), mirrored by slope patterns. They have a narrow range of Eu/Eu* ratios (Zanjan quartz-monzodiorite: 0.67–0.84; Zaker quartz-monzonite: 0.47–0.74; Morvarid monzogranite: 0.84–1.33; Kuh-Tabar quartz-monzodiorite: 0.7–0.81). The monzogranite samples show lower REE abundances and higher contents of P than Zanjan and Zaker samples (Fig. 9).

6.2. Zircon geochronology

Zircon crystals from the Tarom plutonic complex were selected for age determination using the LA-ICP-MS technique. Rocks found in the Tarom complex included the Zanjan quartz-monzodiorite pluton

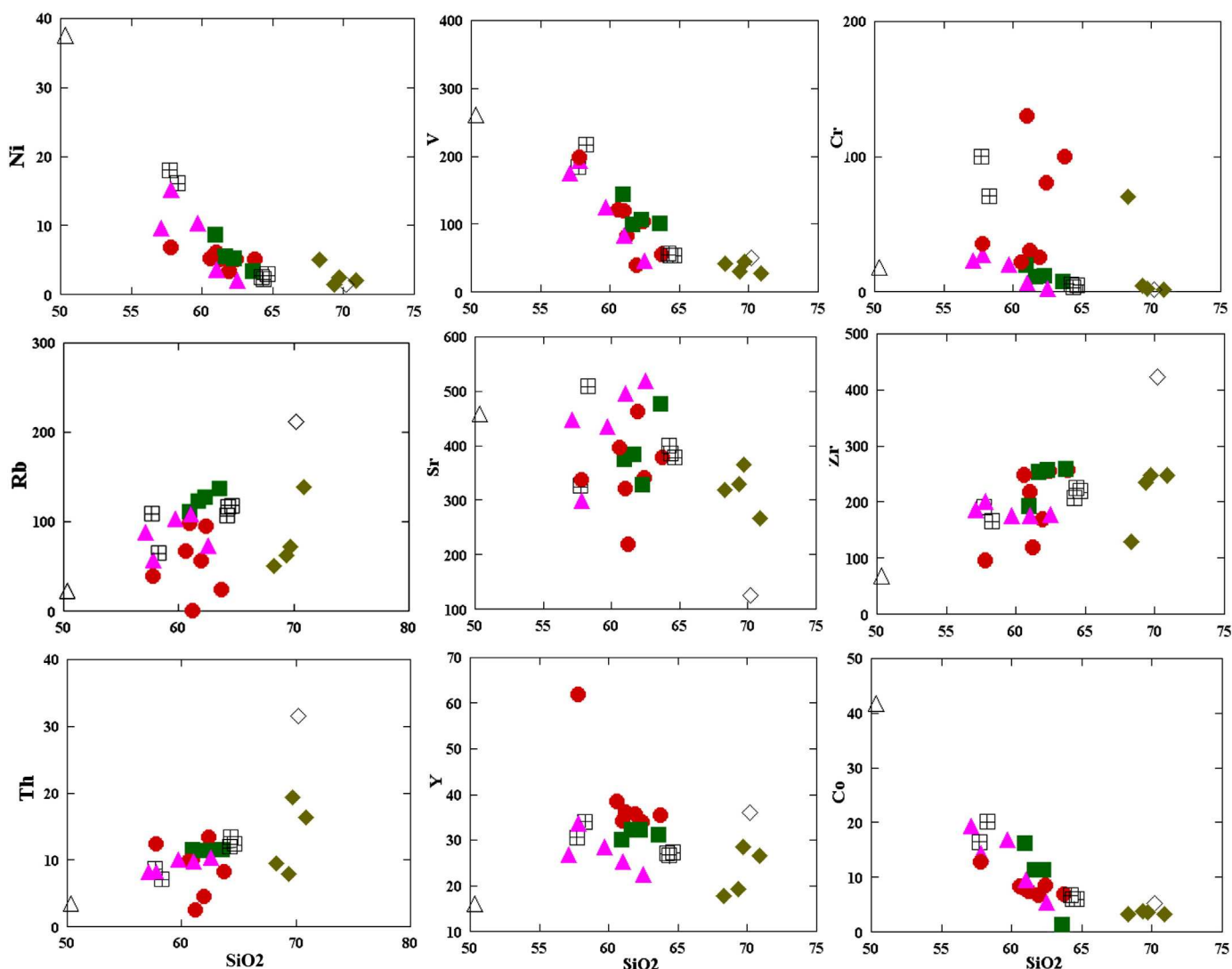


Fig. 7. Selected trace elements vs. SiO₂ (wt.%) contents for the Tarom complex.

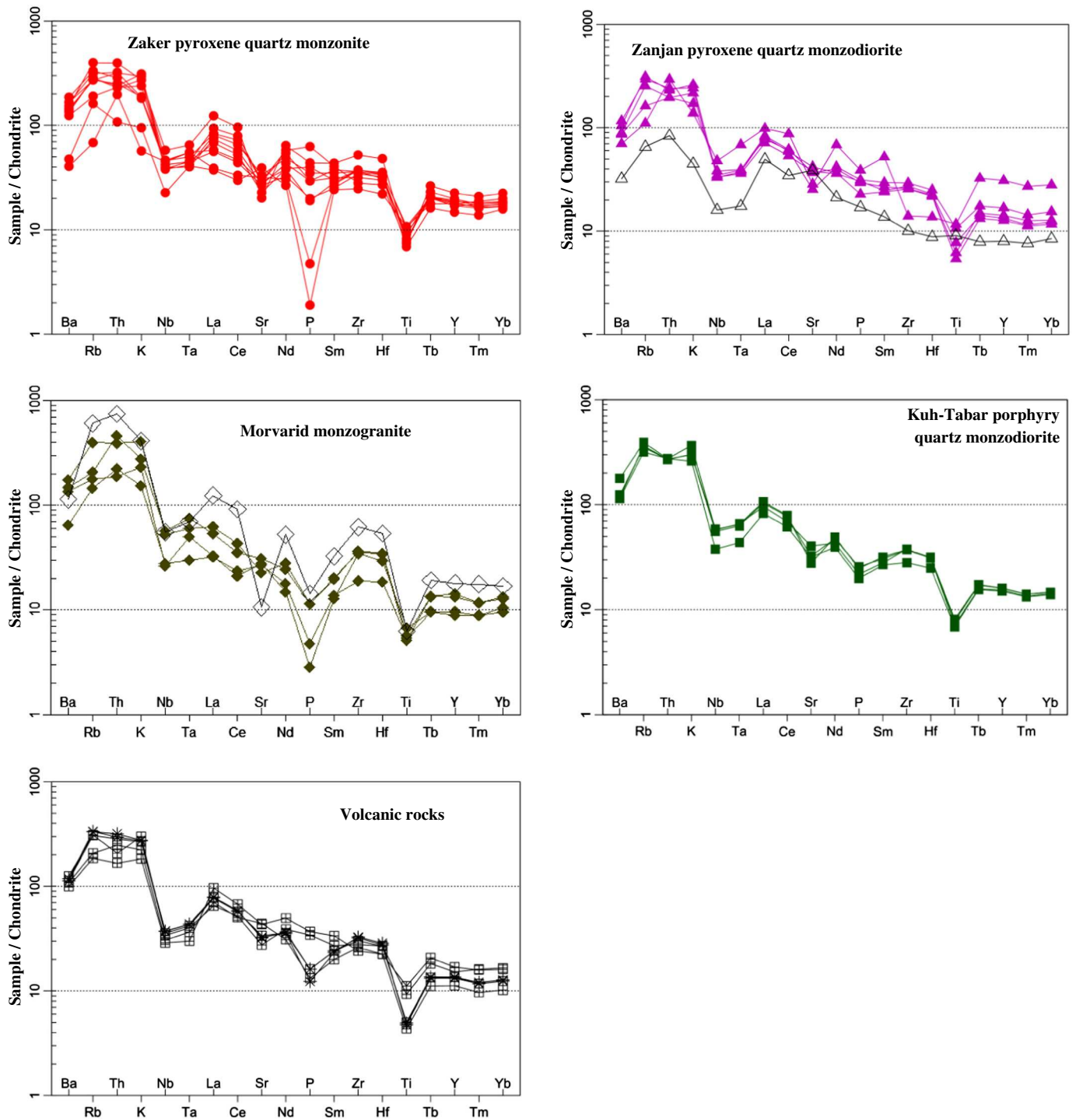


Fig. 8. Chondrite normalized diagrams (Thompson, 1982) showing the main rock units of the Taron.

(N-T19 and N-T52), the Zaker quartz-monzonite pluton (N-SD19) and the Morvarid monzogranite pluton (N-M16) (Table 5). The locations of the analyzed samples are shown in Fig. 2. Unfortunately, there were no sufficient zircon grains in the Kuh-Tabar rocks for successful U–Pb dating. However, field relationships and petrographic studies show a younger age for the Kuh-Tabar pluton than for the Zanjan and Morvarid plutons (see Section 4.4).

The studied zircons in the quartz-monzodiorite rocks have a subhedral to euhedral shape (prismatic) and a light pink color. Zircons of the quartz-monzonite are white to colorless with a lower length/

width aspect ratio than those of the quartz-monzodiorite. Zircons from the monzogranite are pink, long to short prismatic and wider than zircons in the quartz-monzonite (Fig. 10a, b, c, d).

Zircons from the Zanjan intrusion are very low in U abundances and yield concordant ages of 37.4 ± 1.6 Ma and 42.56 ± 0.62 Ma (Fig. 11a, b). Ages of 41.1 ± 1.9 Ma and 39.7 ± 1.2 Ma were obtained for the Zaker quartz-monzonite and Morvarid monzogranite (Fig. 11c, d). Some inherited zircons with very old apparent ages (more than 4 Ga) occur in these intrusive rocks which may be related to U loss in the inherited zircon as a result of temperature effects.

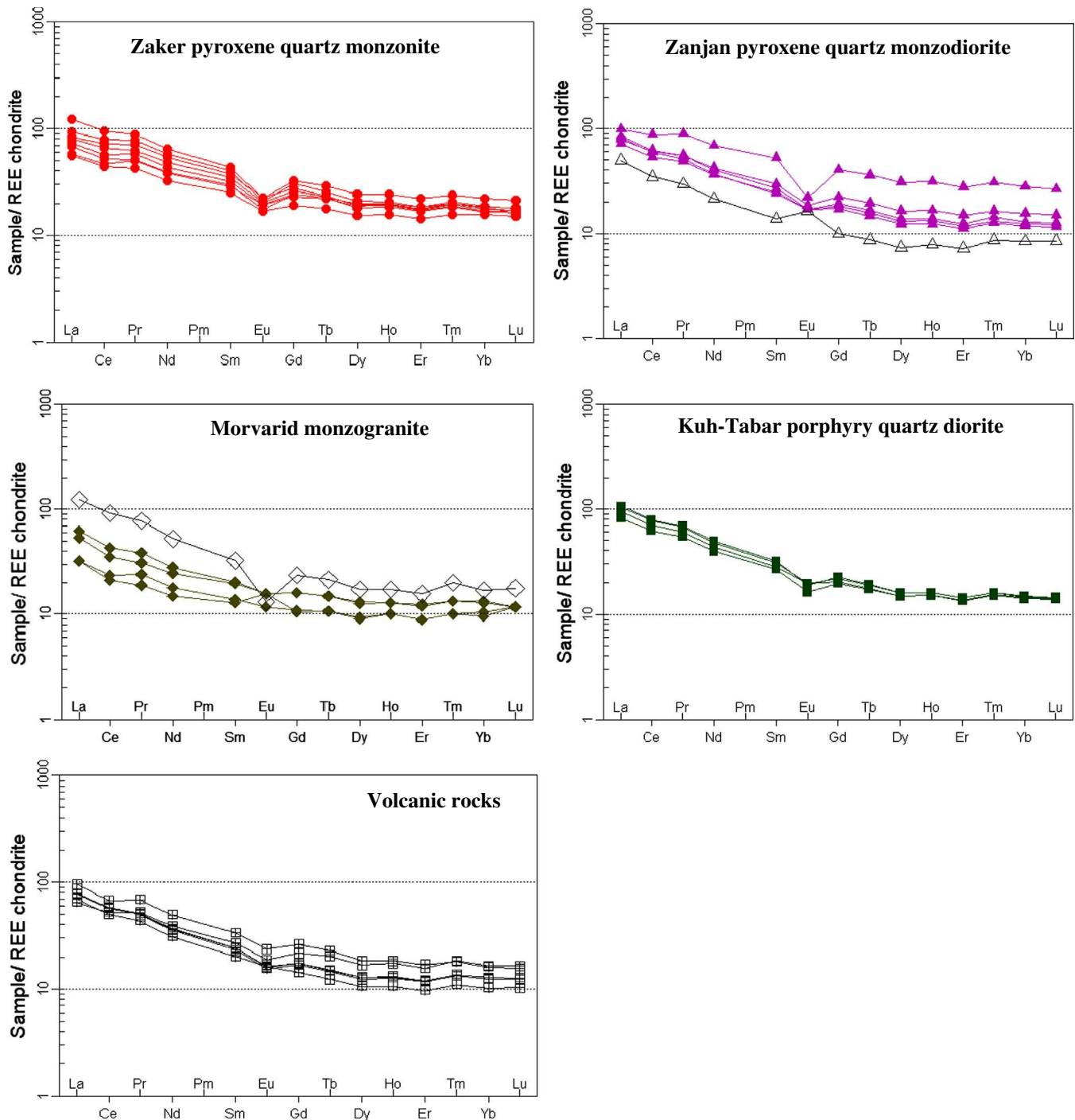


Fig. 9. REE chondrite normalized diagrams (Nakamura, 1974) for the Tarom complex.

6.3. Sr–Nd–Pb isotopes

Neodymium, Sr and Pb isotopic compositions from six samples were studied from the Tarom plutonic rocks (Tables 6 and 7). In order to reveal their source properties, we plotted the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (expressed as $\epsilon_{\text{Nd}}(40 \text{ Ma})$ values) against I_{Sr} . Initial $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were calculated to their average age (40 Ma) according to the results obtained from zircon dating (see Section 5.2). All rocks plot to the right of the mantle array in Fig. 12a and show more variation in $\epsilon_{\text{Nd}}(40 \text{ Ma})$ than in $I_{\text{Sr}}(40 \text{ Ma})$.

All measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios fall within the range of 0.70450–0.70551, and the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are between 0.51231 and 0.51278. The Tarom

intrusives fall within a narrow range for initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.70454–0.70551), whereas $\epsilon_{\text{Nd}}(40 \text{ Ma})$ ratios (–4.2 to +3.4 and –5.7 for an enclave) are variable.

It is apparent that all the samples show some differences in isotopic ratios, which can be related to the crustal assimilation or inherited from the source (Fig. 12a). All rock types exhibit relatively low initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. The depleted mantle Nd model ages (T_{DM}) of all analyzed samples vary in a limited range of 0.7 to 1.3 Ga (Table 6).

The quartz-diorite enclave in the Zanjan pluton shows higher I_{Sr} (0.70505) and lower Nd isotopes ($\epsilon_{\text{Nd}}(42 \text{ Ma}) = -5.72$) relative to the respective host rock. The Nd model age of the enclave ($T_{\text{DM}} = 1.5 \text{ Ga}$) is also slightly older than that of the host rock.

Table 5

U–Pb zircon analytical data of ZMZ plutons in the Tarom area.

Spot	²⁰⁷ Pb/ ²⁰⁶ Pb	1s	²⁰⁷ Pb/ ²³⁵ U	1s	²⁰⁶ Pb/ ²³⁸ U	1s	²⁰⁸ Pb/ ²³² Th	1s	²⁰⁷ Pb/ ²⁰⁶ Pb	1s	²⁰⁷ Pb/ ²³⁵ U	1s	²⁰⁶ Pb/ ²³⁸ U	1s	²⁰⁸ Pb/ ²³² Th	1s
N-SD19-01 ^a	0.04831	0.00206	0.04049	0.00132	0.00608	0.00009	0.00197	0.00004	114.5	97.75	40.3	1.29	39.1	0.59	39.8	0.78
N-SD19-02	0.06487	0.00237	0.05521	0.00134	0.00617	0.00009	0.0022	0.00003	770	75.02	54.6	1.29	39.7	0.58	44.5	0.66
N-SD19-03	0.05925	0.00214	0.05434	0.00129	0.00665	0.0001	0.00216	0.00003	576.2	76.65	53.7	1.24	42.7	0.62	43.5	0.63
N-SD19-04	0.06683	0.00247	0.05564	0.00139	0.00604	0.00009	0.00224	0.00004	832.4	75.35	55	1.33	38.8	0.57	45.3	0.77
N-SD19-05	0.08529	0.00578	0.07305	0.00441	0.00621	0.00013	0.00352	0.00014	1322.1	125.96	71.6	4.17	39.9	0.83	71	2.75
N-SD19-06	0.08174	0.00286	0.07046	0.00154	0.00625	0.00009	0.00251	0.00004	1239.4	66.97	69.1	1.46	40.2	0.58	50.7	0.74
N-SD19-07	0.06218	0.00263	0.05329	0.0017	0.00621	0.0001	0.0019	0.00003	680.2	87.83	52.7	1.64	39.9	0.61	38.4	0.66
N-SD19-08	0.05712	0.00325	0.0454	0.00222	0.00576	0.0001	0.00185	0.00005	495.7	121.12	45.1	2.16	37	0.65	37.3	0.96
N-SD19-09	0.05219	0.00196	0.04741	0.00122	0.00659	0.0001	0.00201	0.00003	293.7	83.28	47	1.18	42.3	0.61	40.6	0.64
N-SD19-10	0.1135	0.00774	0.10862	0.00656	0.00694	0.00015	0.00198	0.00003	1856.2	118.44	104.7	6.01	44.6	0.98	40	0.66
N-SD19-11 ^a	0.04821	0.00168	0.04331	0.00095	0.00651	0.00009	0.00196	0.00003	109.6	80.29	43	0.92	41.9	0.59	39.7	0.54
N-SD19-12	0.05491	0.00253	0.05018	0.00184	0.00663	0.0001	0.00211	0.00005	408.7	99.67	49.7	1.78	42.6	0.67	42.5	0.99
N-SD19-13	0.09291	0.00326	0.09034	0.00198	0.00705	0.0001	0.00293	0.00004	1485.9	65.12	87.8	1.84	45.3	0.66	59	0.86
N-SD19-14	0.05255	0.00207	0.04756	0.00134	0.00656	0.0001	0.00224	0.00004	309.3	87.3	47.2	1.3	42.2	0.62	45.2	0.86
N-SD19-15	0.43525	0.03743	0.93652	0.06551	0.0156	0.00073	0.06364	0.00335	4035.6	122.64	671.1	34.35	99.8	4.66	1247	63.7
N-SD19-16	0.11826	0.00419	0.1109	0.00247	0.0068	0.0001	0.00335	0.00005	1930.1	62.08	106.8	2.26	43.7	0.65	67.7	1.02
N-SD19-17 ^a	0.0473	0.00191	0.04044	0.0012	0.0062	0.00009	0.00182	0.00003	63.7	94.08	40.3	1.17	39.8	0.59	36.8	0.61
N-SD19-18	0.36794	0.02053	0.55245	0.02368	0.01089	0.00031	0.01484	0.00041	3783	82.09	446.6	15.49	69.8	1.96	297.8	8.26
N-SD19-19 ^a	0.05093	0.00184	0.04581	0.00109	0.00652	0.00009	0.00194	0.00003	237.8	81.32	45.5	1.06	41.9	0.6	39.2	0.55
N-SD19-20 ^a	0.04712	0.00181	0.04324	0.00117	0.00665	0.0001	0.00205	0.00003	54.9	89.8	43	1.14	42.8	0.62	41.3	0.63
N-SD19-21	0.358	0.02596	0.43663	0.02575	0.00884	0.00032	0.00216	0.00005	3741.4	106.03	367.9	18.2	56.8	2.06	43.7	0.95
N-SD19-22	0.05933	0.00215	0.04905	0.00117	0.006	0.00009	0.002	0.00003	579.3	76.83	48.6	1.13	38.5	0.56	40.4	0.61
N-SD19-23	0.63203	0.03898	2.40201	0.11044	0.02756	0.00104	0.10143	0.00348	4583.4	86.43	1243.2	32.96	175.3	6.5	1952.7	63.94
N-M-16-01	0.24585	0.01825	0.28624	0.01796	0.00844	0.00028	0.01034	0.0004	3158.2	113.12	255.6	14.18	54.2	1.81	208	8.09
N-M-16-02	0.0798	0.00994	0.07327	0.00861	0.00666	0.00024	0.00288	0.00021	1192.1	227.86	71.8	8.14	42.8	1.51	58.2	4.17
N-M-16-03	0.17754	0.02176	0.20413	0.02264	0.00834	0.0004	0.00911	0.00062	2630.1	190.42	188.6	19.09	53.5	2.58	183.2	12.41
N-M-16-04	0.12746	0.01889	0.11592	0.01599	0.00659	0.00033	0.00502	0.0004	2063.3	240.44	111.4	14.55	42.4	2.12	101.3	8.08
N-M-16-05	0.10184	0.01293	0.10202	0.01209	0.00726	0.00029	0.00393	0.00028	1657.8	218.32	98.6	11.14	46.7	1.88	79.3	5.73
N-M-16-06	0.08346	0.01245	0.07837	0.0111	0.00681	0.00028	0.00329	0.00027	1279.9	266.18	76.6	10.45	43.7	1.82	66.3	5.46
N-M-16-07	0.08213	0.01263	0.07528	0.01099	0.00665	0.00029	0.00397	0.0003	1248.7	274.59	73.7	10.38	42.7	1.85	80.1	6.12
N-M-16-08	0.07286	0.00876	0.06141	0.007	0.00611	0.00019	0.00262	0.00014	1010.2	226.43	60.5	6.69	39.3	1.23	52.9	2.91
N-M-16-09	0.08573	0.01418	0.08242	0.01298	0.00697	0.00032	0.00383	0.00034	1332	290.35	80.4	12.18	44.8	2.06	77.3	6.81
N-M-16-10	0.2207	0.02567	0.22943	0.02367	0.00754	0.00038	0.01211	0.00073	2985.8	175.81	209.7	19.55	48.4	2.41	243.2	14.55
N-M-16-11	0.11137	0.0145	0.09763	0.01189	0.00636	0.00026	0.00297	0.00021	1821.8	219.26	94.6	11	40.8	1.66	60	4.23
N-M-16-12	0.13999	0.00931	0.13167	0.00755	0.00682	0.00017	0.00463	0.00016	2227	110.83	125.6	6.77	43.8	1.08	93.4	3.28
N-M-16-13	0.10146	0.01083	0.08711	0.0086	0.00623	0.00021	0.00327	0.00021	1650.9	185.86	84.8	8.04	40	1.35	66.1	4.24
N-M-16-14	0.35252	0.01737	0.48989	0.01781	0.01008	0.00024	0.01921	0.00048	3717.9	73.04	404.8	12.14	64.6	1.56	384.5	9.5
N-M-16-15	0.06473	0.00609	0.0568	0.00497	0.00636	0.00016	0.00277	0.00014	765.7	186.5	56.1	4.78	40.9	1.04	55.9	2.78
N-M-16-16	0.20332	0.01833	0.18715	0.0148	0.00667	0.00025	0.00591	0.00029	2853	139.66	174.2	12.65	42.9	1.59	119.1	5.79
N-M-16-17	0.25193	0.02002	0.29635	0.01998	0.00853	0.00031	0.0107	0.00046	3196.9	120.39	263.5	15.65	54.8	1.96	215.2	9.14
N-M-16-18	0.10802	0.01413	0.10983	0.0134	0.00737	0.00031	0.00494	0.00036	1766.3	221.66	105.8	12.26	47.4	1.98	99.6	7.19
N-M-16-19	0.05458	0.01216	0.04661	0.01008	0.00619	0.00031	0.00283	0.00028	394.9	434.75	46.3	9.78	39.8	1.97	57.2	5.6
N-M-16-20	0.07466	0.01233	0.06283	0.00992	0.0061	0.00026	0.00246	0.00019	1059	301.35	61.9	9.48	39.2	1.69	49.7	3.83
N-M-16-21	0.05705	0.00851	0.04849	0.00695	0.00616	0.00021	0.00248	0.00017	493	299.47	48.1	6.73	39.6	1.36	50.1	3.49
N-M-16-22	0.51088	0.02978	0.97282	0.04214	0.01381	0.00045	0.02609	0.00079	4272.9	83.19	689.9	21.69	88.4	2.85	520.7	15.52
N-M-16-23	0.0932	0.01392	0.08236	0.01164	0.00641	0.00028	0.00447	0.00033	1491.9	259.09	80.4	10.92	41.2	1.76	90.2	6.7
N-M-16-24	0.38231	0.0228	0.51242	0.02362	0.00972	0.00029	0.01739	0.00055	3840.9	87.24	420.1	15.86	62.4	1.88	348.4	10.83
N-M-16-25	0.07556	0.014	0.07093	0.01258	0.00681	0.00034	0.00308	0.00028	1083.4	332.59	69.6	11.92	43.7	2.15	62.3	5.71
N-M-16-26	0.35477	0.02844	0.47093	0.03094	0.00963	0.00039	0.01391	0.00059	3727.6	116.86	391.8	21.36	61.8	2.49	279.1	11.76
N-M-16-27	0.25049	0.02416	0.2357	0.01968	0.00683	0.00029	0.0084	0.00043	3187.9	144.91	214.9	16.17	43.8	1.88	169.1	8.65
N-M-16-28	0.09379	0.01055	0.08547	0.00895	0.00661	0.00022	0.00334	0.00021	1503.9	198.77	83.3	8.38	42.5	1.43	67.4	4.32
N-M-16-29	0.08675	0.0146	0.07483	0.01198	0.00626	0.0003	0.00288	0.00027	1354.9	294.07	73.3	11.32	40.2	1.89	58.1	5.49
N-M-16-30	0.09882	0.01044	0.0934	0.00914	0.00686	0.00022	0.00436	0.00025	1602	185.1	90.7	8.49	44	1.42	87.9	5.06
N-T-19-01	0.20154	0.00771	0.19183	0.00459	0.0069	0.00011	0.00394	0.00007	2838.7	61.05	178.2	3.91	44.3	0.69	79.6	1.31
N-T-19-02	0.07836	0.00301	0.06334	0.00159	0.00586	0.00008	0.00368	0.00007	1156.1	74.4	62.4	1.51	37.7	0.53	74.3	1.48
N-T-19-03	0.05919	0.00334	0.04778	0.00227	0.00585	0.0001	0.0025	0.00006	573.9	118.15	47.4	2.2	37.6	0.62	50.5	1.16
N-T-19-04	0.08795	0.00454	0.07719	0.00299	0.00593	0.0001	0.00222	0.00005	1381.4	96.02	70.5	2.84	38.1	0.66	44.8	0.95
N-T-19-05	0.11085	0.00451	0.09459	0.00265	0.00619	0.00009	0.00605	0.00013	1813.4	72.06	91.8	2.46	39.8	0.6	122	2.63
N-T-19-06	0.13155	0.00523	0.11997	0.00321	0.00661	0.0001	0.00285	0.00005	2118.7	68.12	115	2.91	42.5	0.65	57.6	0.96
N-T-19-07	0.16606	0.00626	0.17135	0.00407	0.00748	0.00011	0.00437	0.00007	2518.3	62	160.6	3.53	48	0.73	88.2	1.42
N-T-19-08	0.21283	0.01498	0.20208	0.01201	0.00688	0.0002	0.00487	0.00017	2927.2	109.5	186.9	10.14	44.2	1.27	98.2	3.45
N-T-19-09	0.15801	0.00588	0.18474	0.00429	0.00847	0.00013	0.00869	0.00015	2434.5	61.67	172.1	3.67	54.4	0.81	174.8	3.08
N-T-19-10	0.0813	0.00345	0.07161	0.00221	0.00638	0.0001	0.00236	0.00004	1228.7	80.97	70.2	2.09	41	0.62	47.7	0.89
N-T-19-11 ^a	0.05106	0.00217	0.04396	0.00138	0.00624	0										

Table 5 (continued)

Spot	²⁰⁷ Pb/ ²⁰⁶ Pb	1s	²⁰⁷ Pb/ ²³⁵ U	1s	²⁰⁶ Pb/ ²³⁸ U	1s	²⁰⁸ Pb/ ²³² Th	1s	²⁰⁷ Pb/ ²⁰⁶ Pb	1s	²⁰⁷ Pb/ ²³⁵ U	1s	²⁰⁶ Pb/ ²³⁸ U	1s	²⁰⁸ Pb/ ²³² Th	1s
N-T-19-21	0.06557	0.00289	0.05364	0.0018	0.00593	0.00009	0.002	0.00003	792.7	89.71	53.1	1.74	38.1	0.59	40.4	0.68
N-T-19-22	0.11892	0.00626	0.10773	0.00463	0.00657	0.00013	0.00262	0.00006	1940.1	91.26	103.9	4.24	42.2	0.81	52.8	1.17
N-T-19-23	0.05706	0.00259	0.04939	0.00175	0.00628	0.0001	0.00206	0.00004	493.1	97.6	48.9	1.7	40.3	0.62	41.5	0.72
N-T-19-24	0.18327	0.00653	0.19233	0.00421	0.00761	0.00012	0.00355	0.00005	2682.7	57.73	178.6	3.59	48.9	0.74	71.7	1.09
N-T-19-25	0.10804	0.0044	0.09941	0.00291	0.00667	0.00011	0.00369	0.00007	1766.5	72.59	96.2	2.69	42.9	0.68	74.5	1.47
N-T-19-26 ^a	0.04969	0.00249	0.0409	0.00169	0.00597	0.0001	0.00186	0.00003	180.3	112.84	40.7	1.65	38.4	0.61	37.6	0.69
N-T-19-27	0.05921	0.00384	0.05439	0.00314	0.00666	0.00013	0.0019	0.00004	574.8	135.25	53.8	3.02	42.8	0.81	38.4	0.9
N-T-19-28	0.10282	0.00394	0.09159	0.00242	0.00646	0.0001	0.0022	0.00004	1675.6	69.21	89	2.25	41.5	0.64	44.5	0.73
N-T-19-29	0.10451	0.00489	0.08925	0.00328	0.00619	0.00011	0.00141	0.00003	1705.6	83.67	86.8	3.06	39.8	0.69	28.5	0.57
N-T-19-30	0.07644	0.00365	0.06528	0.0025	0.0062	0.0001	0.00233	0.00005	1106.5	92.47	64.2	2.38	39.8	0.66	47	0.97
N-T-52-01	0.05558	0.00204	0.04849	0.00122	0.00632	0.00009	0.00204	0.00003	435.3	79.59	48.1	1.18	40.6	0.57	41.1	0.57
N-T-52-02	0.06271	0.00227	0.05204	0.00127	0.00601	0.00008	0.00193	0.00003	698.5	75.29	51.5	1.22	38.7	0.54	39	0.55
N-T-52-03	0.0699	0.00261	0.05953	0.00154	0.00617	0.00009	0.002	0.00003	925.4	74.94	58.7	1.48	39.7	0.57	40.3	0.58
N-T-52-04	0.07184	0.00304	0.0574	0.00185	0.00579	0.00009	0.00195	0.00003	981.3	83.83	56.7	1.77	37.2	0.57	39.4	0.65
N-T-52-05	0.05807	0.00241	0.05379	0.00169	0.00672	0.0001	0.00221	0.00003	532	88.76	53.2	1.63	43.1	0.64	44.7	0.69
N-T-52-06	0.08966	0.00336	0.073	0.00189	0.0059	0.00009	0.00216	0.00003	1418.2	70.12	71.5	1.79	37.9	0.56	43.6	0.66
N-T-52-07	0.05255	0.00223	0.04547	0.00148	0.00627	0.00009	0.00194	0.00003	309.5	93.7	45.1	1.44	40.3	0.6	39.2	0.62
N-T-52-08	0.06219	0.00233	0.05064	0.00133	0.0059	0.00009	0.0019	0.00003	680.7	78.16	50.2	1.28	38	0.55	38.4	0.55
N-T-52-09	0.06172	0.00233	0.05517	0.00146	0.00648	0.00009	0.00212	0.00003	664.3	78.83	54.5	1.41	41.7	0.6	42.7	0.61
N-T-52-10	0.07728	0.00331	0.06114	0.002	0.00574	0.00009	0.0021	0.00003	1128.4	82.92	60.3	1.91	36.9	0.58	42.5	0.71
N-T-52-11	0.05216	0.00192	0.04511	0.00115	0.00627	0.00009	0.00031	0.00002	292.6	81.91	44.8	1.11	40.3	0.57	6.2	0.49
N-T-52-12 ^a	0.04874	0.0019	0.04497	0.00127	0.00669	0.0001	0.00215	0.00003	135.1	89.11	44.7	1.24	43	0.62	43.4	0.64
N-T-52-13	0.16217	0.00711	0.1613	0.00533	0.00721	0.00013	0.00415	0.00008	2478.5	72.16	151.8	4.66	46.3	0.84	83.6	1.58
N-T-52-14 ^a	0.04623	0.00167	0.0421	0.00103	0.00661	0.00009	0.00213	0.00003	9.6	83.59	41.9	1	42.4	0.6	43	0.6
N-T-52-15	0.07737	0.00261	0.06711	0.00139	0.00629	0.00009	0.00224	0.00003	1130.7	65.83	66	1.33	40.4	0.57	45.3	0.61
N-T-52-16	0.4731	0.01958	1.06629	0.02953	0.01635	0.00036	0.01606	0.00031	4159.5	59.97	736.9	14.51	104.5	2.26	322	6.12
N-T-52-17	0.05948	0.00264	0.05365	0.00188	0.00654	0.0001	0.00218	0.00004	584.8	93.54	53.1	1.81	42	0.65	44.1	0.73
N-T-52-18	0.05078	0.00251	0.04552	0.00187	0.0065	0.0001	0.00221	0.00004	230.8	110.22	45.2	1.81	41.8	0.67	44.5	0.88
N-T-52-19	0.05273	0.00201	0.04885	0.00133	0.00672	0.0001	0.00211	0.00003	317.1	84.64	48.4	1.29	43.2	0.63	42.6	0.61
N-T-52-20	0.13381	0.0052	0.11204	0.00306	0.00607	0.0001	0.00272	0.00004	2148.5	66.39	107.8	2.79	39	0.63	55	0.89
N-T-52-21 ^a	0.04689	0.0017	0.04359	0.00107	0.00675	0.0001	0.00208	0.00003	43.4	84.62	43.3	1.04	43.3	0.62	41.9	0.57
N-T-52-22 ^a	0.04631	0.00166	0.04288	0.00103	0.00672	0.0001	0.00212	0.00003	13.6	82.8	42.6	1	43.2	0.61	42.9	0.6
N-T-52-23 ^a	0.0461	0.00157	0.04178	0.00089	0.00657	0.00009	0.00202	0.00003	3.1	78.98	41.6	0.87	42.2	0.59	40.8	0.53
N-T-52-24 ^a	0.04885	0.00193	0.0438	0.00127	0.00651	0.0001	0.00207	0.00003	140.9	90.39	43.5	1.24	41.8	0.62	41.9	0.62
N-T-52-25 ^a	0.04857	0.00203	0.04524	0.00144	0.00676	0.0001	0.00208	0.00003	127.3	95.28	44.9	1.4	43.4	0.65	41.9	0.64
N-T-52-26	0.09541	0.00329	0.1001	0.00218	0.00761	0.00011	0.00293	0.00004	1536.1	63.56	96.9	2.01	48.9	0.72	59.2	0.84
N-T-52-27 ^a	0.04704	0.00174	0.04173	0.00107	0.00644	0.00009	0.00203	0.00003	50.9	86.33	41.5	1.04	41.4	0.6	41	0.57
N-T-52-28	0.05597	0.00288	0.05099	0.0022	0.00661	0.00011	0.0021	0.00004	450.7	110.55	50.5	2.13	42.5	0.71	42.4	0.8
N-T-52-29	0.05326	0.00275	0.04561	0.00199	0.00622	0.0001	0.00215	0.00004	339.8	112.77	45.3	1.93	39.9	0.66	43.5	0.79
N-T-52-30	0.05172	0.00242	0.04546	0.00172	0.00638	0.0001	0.00221	0.00004	272.8	103.6	45.1	1.67	41	0.65	44.5	0.79

^a denotes concordant spots used to calculate the corresponding mean age.

All samples from the study area plot near the Urumieh-Dokhtar Eocene arc rocks from the Khalkhab (Rezaei-Kahkhaei et al., 2011), back arc shoshonitic rocks in the Central Iran (Ahmadian et al., 2009), potassic syenitic pluton in the Scottish Caledonia (Fowler, 1992) and adakitic post-collisional plutonism of the eastern Pontides in Turkey (Topuz et al., 2005) and show isotopic ratios similar to the ~40-km-thick Andean arc crust (Haschke et al., 2002; Petford and Atherton, 1996) (Fig. 12b).

The Tarom samples are relatively homogeneous in Pb isotopic compositions of ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb isotopic ratios. Lead isotopic ratios of the samples are ²⁰⁶Pb/²⁰⁴Pb = 18.52–18.86, ²⁰⁷Pb/²⁰⁴Pb = 15.57–15.72, and ²⁰⁸Pb/²⁰⁴Pb = 38.47–39.08 (Table 7). The average Pb isotopic ratios of the ZZM rocks are (²⁰⁶Pb/²⁰⁴Pb) = 18.67, (²⁰⁷Pb/²⁰⁴Pb) = 15.62, and (²⁰⁸Pb/²⁰⁴Pb) = 38.74, which is similar to that of the Zanjan enclave [(²⁰⁶Pb/²⁰⁴Pb) = 18.67, (²⁰⁷Pb/²⁰⁴Pb) = 15.63, (²⁰⁸Pb/²⁰⁴Pb) = 38.83]. The Pb isotopic ratios of the Kuh-Tabar sample are (²⁰⁶Pb/²⁰⁴Pb) = 18.69, (²⁰⁷Pb/²⁰⁴Pb) = 15.61, and (²⁰⁸Pb/²⁰⁴Pb) = 38.78.

The Northern Hemisphere Reference Line (NHRL; Hart, 1984) was used in the plots since ²⁰⁸Pb data are more radiogenic than ²⁰⁶Pb. All the samples lie well above the NHRL in conventional Pb isotope diagrams (Fig. 13a, b) and plot between EMI (enriched mantle with intermediate ⁸⁷Sr/⁸⁶Sr, low ¹⁴³Nd/¹⁴⁴Nd and low ²⁰⁶Pb/²⁰⁴Pb) and EMII (enriched mantle with high ⁸⁷Sr/⁸⁶Sr, intermediate ¹⁴³Nd/¹⁴⁴Nd and high ²⁰⁶Pb/²⁰⁴Pb; Zindler and Hart, 1986) but closer to EMII. Pb isotopes

of the samples are clustered in a field close to EMII (Fig. 13a, b) and marine sediments (Fig. 13c).

7. Discussion

7.1. Petrogenetic considerations of the Tarom complex

All intrusions from the Tarom plutonic complex have close U–Pb zircon ages and covariant geochemical compositions indicating that they may have similar petrogenesis and magma source. The main differentiation process in the Tarom complex seems to be fractional crystallization from a unique parental magma (with some assimilation in continental crust; AFC model) or partial melting from a common source. However, most of the geochemical and isotopic evidences point to a partial melting process rather than a fractional crystallization process. For instance, Harker diagrams (Fig. 6) do not show linear trends in most elements (especially in the Zaker pluton) which is expected for fractional crystallization process, although major and trace elements versus SiO₂ diagrams (Figs. 6 and 7) are consistent with fractionation of plagioclase, clinopyroxene, and Fe–Ti oxides during magmatic evolution of each pluton separately. Moreover, REE chondrite-normalized diagrams (Fig. 9) exhibit patterns with very limited variations in HREE but with a higher variation in LREE (for example, the Kuh-Tabar and Zaker plutons). These patterns are typical of a partial melting process or a crystal fractionation process involving mineral phases that strongly

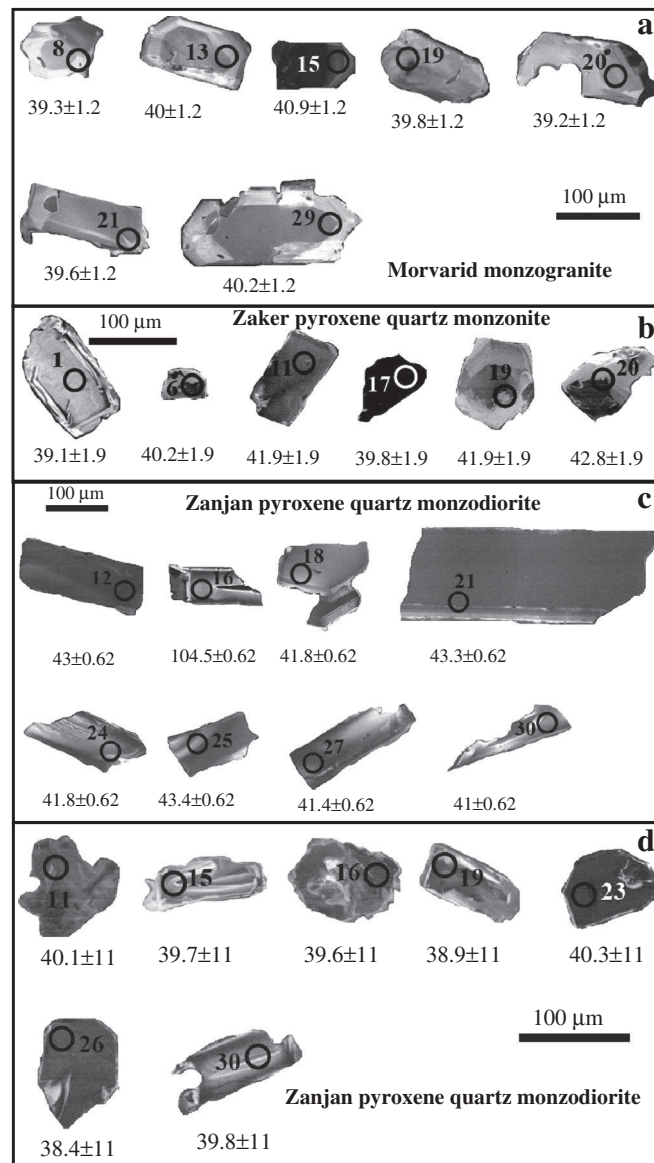


Fig. 10. CL images of some representative zircon crystals used for U–Pb dating by LA-ICP-MS from the (a) Morvarid monzogranite (N-M16); (b) Zaker quartz monzonite (N-SD19); (c) Zanjan quartz monzodiorite (N-T52) and (d) Zanjan quartz monzodiorite (N-T19). Note that some contain inherited cores.

fractionate LREE relative to HREE, such as monazite or allanite (e.g., López-Moro and López-Plaza, 2004; López-Moro et al., 2012). Based on petrographic studies, there is no monazite and allanite in the Tarom monzonitic rocks. According to the mineralogical and geochemical composition, partial melting process played the major role in the differentiation of Zanjan, Zaker and Kuh-Tabar plutons, as is also suggested by the La/Yb vs. La diagram (Fig. 14a). Furthermore, La/Yb ratio is higher in the most differentiated samples (Fig. 14a, Table 4), ruling out the process involving fractionated mineral phases that strongly fractionate LREE relative to HREE.

The Isr ratios of the studied samples exhibit neither negative nor positive correlations with increasing SiO_2 contents, confirming partial melting/fractional crystallization as the major process in their genesis (Fig. 14b). However, a quartz-monzodiorite sample (N-T19; $\epsilon_{Nd(40\text{ Ma})} = -4.2$) shows lower value of $\epsilon_{Nd(40\text{ Ma})}$ than a more evolved sample (e.g., N-M65; $\epsilon_{Nd(40\text{ Ma})} = +2.1$) which is not consistent with AFC process. In addition, there is no evidence for magma mixing. Mafic microgranular enclaves do not occur in great abundance and are only found in some parts of the pluton margins. In addition,

the absence of disequilibrium textures such as antirapakivi, sieve texture, quartz ocelli and needle apatite in the enclaves disproves magma mixing processes in the generation of the rocks.

7.2. Source characteristics

The contrasting Nd–Sr isotopic compositions between the Tarom samples (Table 6) and the mafic lower crust (e.g., Rudnick, 1990) rule out the last one as a potential source of Tarom magmas and dominantly point to mantle as the main source for magma genesis. The Nb/La ratios of the Tarom samples also range from 0.3 to 1.2 suggesting a lithospheric mantle source (Bradshaw and Smith, 1994; Smith et al., 1999). In addition, the chondrite-normalized patterns of studied samples show a significant enrichment of their mantle source. The Ce/Pb ratios (0.75–15) of the samples are considerably different from those of oceanic basalts (~25; Hofmann, 1997), indicating that these rocks are not derived from normal asthenospheric mantle melts. The Tarom samples are characterized by high concentrations of incompatible trace elements (LILE and LREE) and they are pronounced by significant negative Nb, Ta

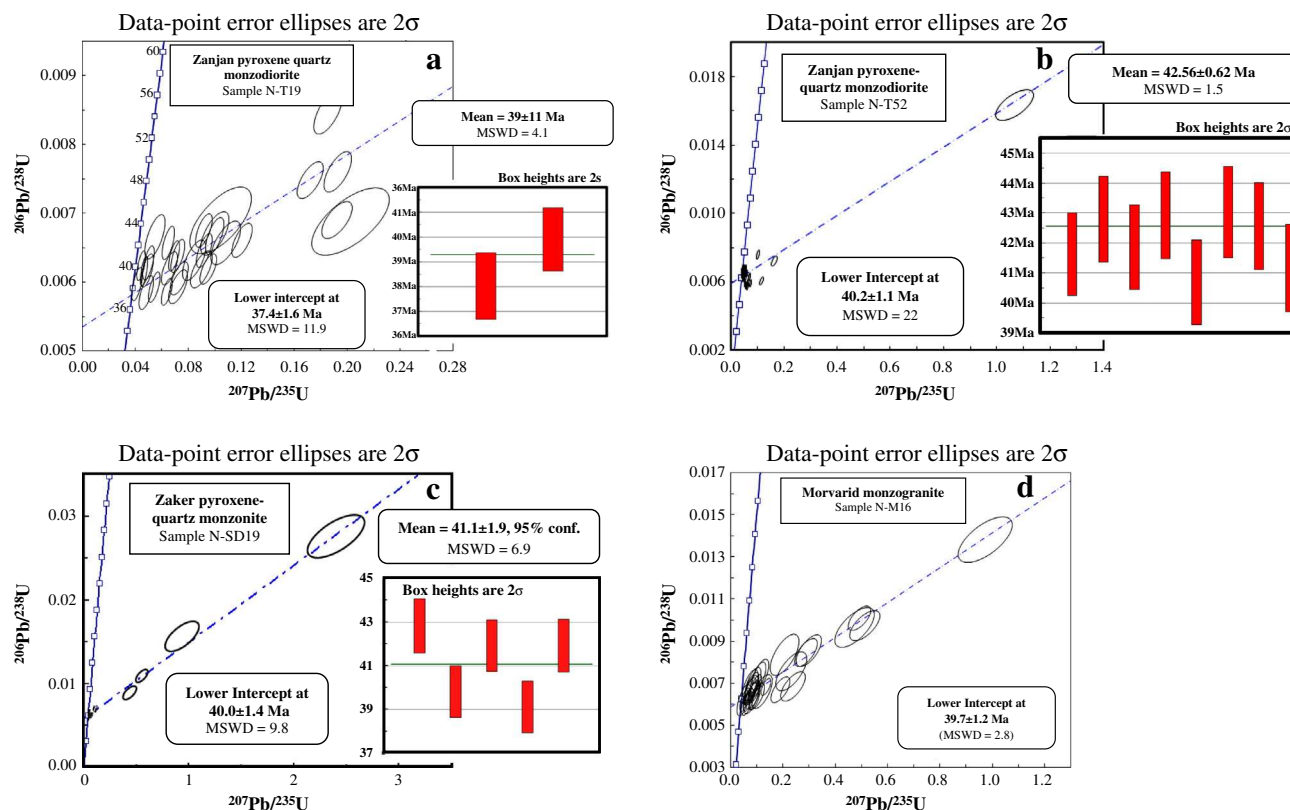


Fig. 11. Zircon U–Pb concordia diagram for representative plutonic rocks of the Tarom intrusive complex; (a) Zanjan quartz monzodiorite (sample N-T19); (b) Zanjan quartz monzodiorite (sample N-T52); (c) Zaker quartz monzonite (sample N-SD19) and (d) Morvarid monzogranite to syenogranite (sample N-M16).

and Ti anomalies (Fig. 8), implying that the samples were not derived from normal MORB or OIB mantle sources, which typically display positive Nb–Ti anomalies in chondrite-normalized trace element diagrams (e.g., Hofmann, 1997).

Isotopic and geochemical data such as I_{Sr} , ϵ_{Nd} , enriched LILE and LREE, show that the source rocks of the Tarom complex must be chemically enriched lithospheric mantle, resembling an EMII-type end-member, which is generally interpreted as the recycled product of terrigenous sediments from an earlier subduction setting (Beccaluva et al., 2004; Zindler and Hart, 1986). These interpretations are well-matched with Pb isotopic ratios of the samples indicating an important role of a chemically enriched mantle source (EMII; Fig. 13a, b) in their generation. High potassic to shoshonitic affinity, K_2O and TiO_2 contents as well as Th/U and Rb/Sr ratios are relatively similar to those of the potassic magmatism in the Eastern Pontides in NE Turkey (Sisdagi pluton)

and the Alborz belt (Shaivar Dagh plutonic complex), where Karsli et al. (2012) and Aghazadeh et al. (2011) took these geochemical features as indicative of sources affected by subduction, including a sedimentary component. Conceição and Green (2004) have performed an approach to the melting of metasomatized peridotite consisting of peridotite with phlogopite and pargasite that yielded similar melts to quartz-monzodiorite and quartz-monzonite rocks from the Tarom complex. However, Morvarid granite displays a remarkable high SiO_2 compared to those of the experimental results of Conceição and Green (2004) which may suggest fractional crystallization from quartz-monzonite to monzogranite as the main process for monzogranite formation.

The difference between element compatibility of phlogopite and amphibole shows which hydrous phase formed in the lithospheric mantle source (Furman and Graham, 1999; Yang et al., 2004). Since the Rb/Sr ratios of granites can reflect the Rb/Sr ratios of their source

Table 6

Sr–Nd isotopic compositions of Tarom magmatic rocks.

Sample	Rock type	Age	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Error	I_{Sr}	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	Error	$\epsilon_{Nd}(0)$	$\epsilon_{Nd}(40\text{ Ma})$	T_{DM1} (Ga)
GN-M65	Morvarid monzogranite	40	0.6	0.70515	0.000003	0.70458	0.1545	0.512731	0.000009	1.8	2.2	1.1
N-M 16	Morvarid monzogranite	40	1.5	0.70566	0.000003	0.70482	0.1399	0.512690	0.000007	1.0	1.6	0.95
GN SD86	Zaker quartz monzonite	40	0.9	0.70540	0.000003	0.70452	0.1359	0.512788	0.000009	2.9	3.4	0.71
N-SD19	Zaker quartz monzonite	40	0.5	0.70503	0.000002	0.70476	0.1396	0.512558	0.000009	−1.6	−1.0	1.2
GN1	Zaker quartz monzonite	40	0.9	0.70533	0.000003	0.70452	0.1497	0.512757	0.000009	2.3	2.7	0.93
GN-G7	Zaker quartz monzonite	40	0.8	0.70526	0.000003	0.70456	0.1365	0.512739	0.000009	2.0	2.5	0.81
GN-T38	Zanjan quartz monzodiorite	40	0.7	0.70492	0.000003	0.70434	0.1308	0.512733	0.000009	1.9	2.4	0.76
N-T 19	Zanjan quartz monzodiorite	40	0.6	0.70490	0.000007	0.70454	0.1271	0.512424	0.000004	−4.2	−3.6	1.3
N-T52	Zanjan quartz monzodiorite	40	0.6	0.70491	0.000005	0.70459	0.1327	0.512661	0.000004	0.5	1.0	0.93
GN-T57	Zanjan quartz monzodiorite	40	0.6	0.70495	0.000003	0.70450	0.1367	0.512731	0.000009	1.8	2.3	0.83
N-D9	Kuh-Tabar quartz monzodiorite	40	1.1	0.70614	0.000009	0.70552	0.1259	0.512389	0.000009	−4.9	−4.3	1.3
N-T52 En	Quartz diorite (enclave)	40	0.1	0.70514	0.000025	0.70505	0.1259	0.512313	0.000007	−6.3	−5.7	1.4

Table 7
Pb isotopic composition of Tarom magmatic rocks.

Samples	Rock type	$^{206}\text{Pb}/^{204}\text{Pb}$	Error	$^{207}\text{Pb}/^{204}\text{Pb}$	Error	$^{208}\text{Pb}/^{204}\text{Pb}$	Error	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
N-SD19	Zaker quartz monzonite	18.7909	0.0036	15.6109	0.0034	38.9996	0.0079	18.65690	15.604626	38.7369
N-M 16	Morvarid monzogranite	18.7503	0.0069	15.5957	0.0057	38.9113	0.0075	18.65669	15.591317	38.69819
N-T52 En	Quartz diorite (Enclave)	18.6918	0.0025	15.6353	0.0023	38.8649	0.0093	18.672910	15.634415	38.83437
N-T 19	Zanjan quartz monzodiorite	18.6029	0.003	15.5744	0.0028	38.5686	0.0056	18.524138	15.570712	38.4731
N-T52	Zanjan quartz monzodiorite	18.9653	0.0017	15.7278	0.0015	39.1397	0.0039	18.861380	15.722934	39.08312
N-D9	Kuh-Tabar quartz monzodiorite	18.7463	0.0025	15.6097	0.00018	38.8404	0.0058	18.697307	15.607406	38.78439

rocks (Kemp and Hawkesworth, 2003), the source of the Tarom complex would have high Rb/Sr ratios, but a low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, likely indicating a mica-bearing source with a short residence time within the mantle. The Ba/Rb (6.63–12.42) and Nb/Th (1.20–1.63) ratios of Tarom samples are relatively low, whereas their Rb/Sr ratios (0.05–0.24) are relatively high, suggesting that the source region contains phlogopite rather than amphibole. Spinel is depleted in REE and Y, while garnet is enriched in HREE and Y, and amphibole is enriched in Middle REE (MREE). $(\text{La}/\text{Yb})_N$ values in the Tarom samples (<7.5) are less than

those in the magma [$(\text{La}/\text{Yb})_N > 20$; e.g., Martin (1987)] which have garnet as the major phase in their source, which precludes the presence of garnet as the major residual phase. Thus, spinel and/or amphibole may be present in residue. However, the slight enrichment in MREE of the

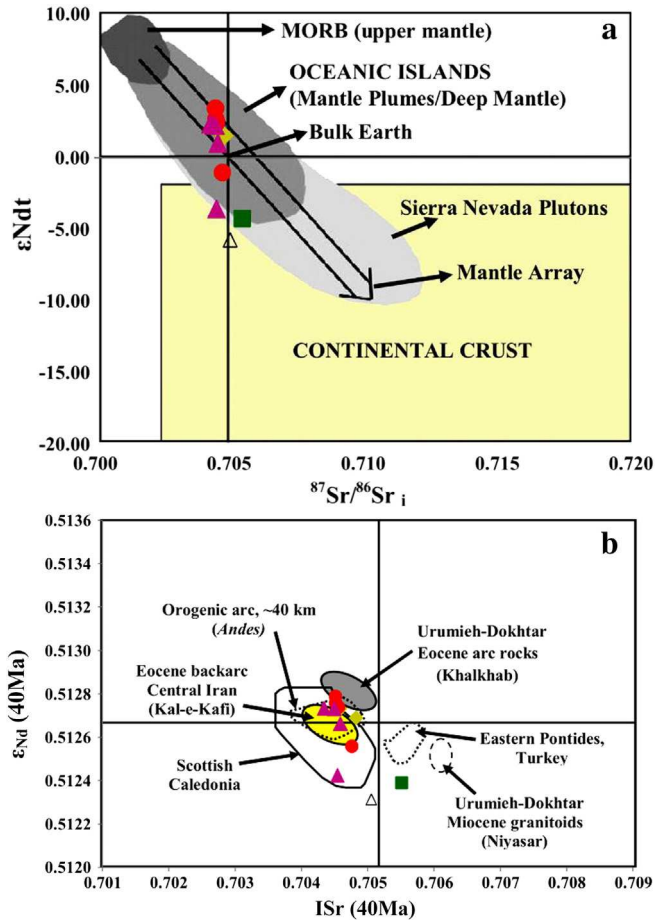


Fig. 12. Nd–Sr isotope compositions of the Tarom complex. Data source are as follows: 400–179 Ma MORB are from Mahoney et al. (1989), Xu et al. (2003), Tribuzio et al. (2004) and Xu and Castillo (2004). (b) Nd and Sr isotopic initial ratios for representative rocks of the Tarom complex. Other plutonic and volcanic rocks from magmatic belts in Iran and adjacent areas are included for comparison. Eocene plutons in the central part of Urumieh-Dokhtar (Ahmadian et al., 2009), Urumieh-Dokhtar Eocene arc rocks from Khalkhab (Rezaei-Kahkhaei et al., 2011), potassic syenitic pluton in the Scottish Caledonia (Fowler, 1992), adakitic post-collisional plutonism eastern Pontides in Turkey (Topuz et al., 2005), Urumieh-Dokhtar Miocene post-collisional granitoids from Niyasar (Honarmand et al., 2013) and show isotope ratios, similar to ~40-km-thick Andean arc crust (Haschke et al., 2002; Petford and Atherton, 1996).

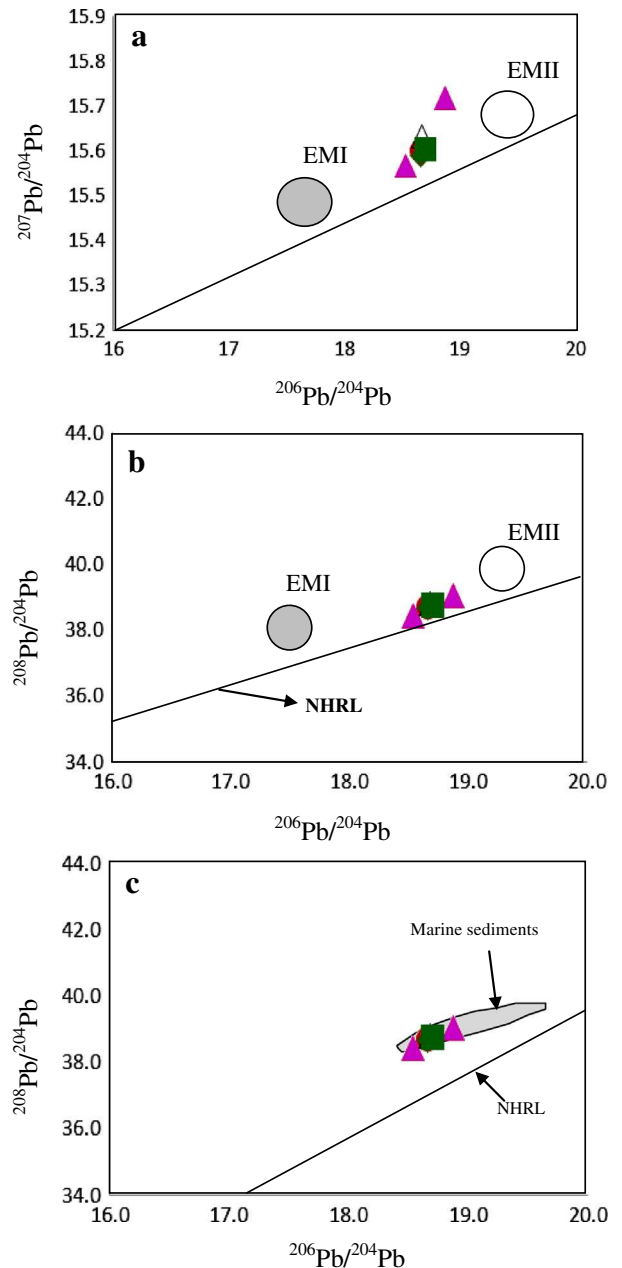


Fig. 13. Plots of (a) $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$; (b) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and (c) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for the selected samples from the Tarom complex (areas of EMI, EMII and marine sediments after Zindler and Hart, 1986).

samples (Fig. 9) suggests that amphibole may not remain stable in the residue after partial melting. Geochemical and isotopic data of this study suggest phlogopite-bearing spinel peridotite as the source materials of the Tarom magmas.

It must be considered that Nd–Sr isotopes (Table 6) of the Tarom plutonic rocks show some variations which rule out a common homogeneous source for the Tarom magmas. Thus, the similar ages of the plutons as well as some variety in Nd–Sr values may be explained by different partial melting degrees of some similar protoliths in a close time. Since, metasomatic influxes could have heterogeneous effects on the pristine mantle, the degree of partial melting of the volatile-enriched portions increased (Viccaro and Cristofolini, 2008). Thus, different partial melting degrees of a mantle re-fertilized by silicate melts/fluids, inheriting from hydrothermally altered oceanic slab stored for a long time in the mantle, could produce batches of melts that result in monzonitic plutons of the Tarom complex.

Volcanic rocks are enriched in LREE and LILE and depleted in HFSE, bearing a weak Eu anomaly in chondrite-normalized trace element patterns (Figs. 8 and 9). Geochemical features of the Tarom volcanic rocks are similar to the plutonic samples which may attribute to a common metasomatized mantle source for the Tarom volcanic and plutonic activity.

7.3. Geodynamic scenario for the Tarom high-potassic plutonism

As it was mentioned earlier, shoshonitic and high-potassic magmatic rocks are commonly formed in arc and post-collision settings (e.g., Foley and Peccerillo, 1992; Turner et al., 1996), but rarely observed in within-plate environments (Bonin, 2004; Müller and Groves, 1995). Post-

collisional, potassic magmatism with subduction-related geochemical signatures is a widespread feature of orogenic collisional orogens. The enrichments in LILE and LREE with negative Nb–Ti anomalies are characteristics of subduction-related magmas and are commonly related to a mantle source, which has previously been enriched in LILEs over HFSEs by metasomatic activity of fluids derived from the subducted slab or sediments (e.g., Cameron et al., 2003; Hawkesworth et al., 1997; Pearce, 1983). A variety of models have been proposed to interpret such a signature in the post-collisional setting. López-Moro and López-Plaza (2004) suggested a hybrid source of peridotite, amphibolite and metapelite at the crust–mantle boundary while Li et al. (2000) considered the mixing of magmas to originate in the asthenosphere and enriched lithospheric mantle. Taking all evidence into consideration, we favor an enrichment of lithospheric mantle by earlier subduction events in a post-collision stage, as proposed by many authors (Aydin et al., 2008; Eklund et al., 1998; Janous'ek et al., 2000; Liu et al., 2002; Turner et al., 1996). In the Tarom area, however, an active subduction system during Eocene is not reported. Therefore, the generation of the high-potassic to shoshonitic samples seems unlikely to have occurred in an active subduction setting. Additionally, the high LILE/HFSE ratios and negative Nb and Ti anomalies can also be characteristics of magmas that originated from an enriched subcontinental lithospheric mantle caused by earlier subduction events. The depletion in Nb, Ta and Ti can also be explained by retention in Ti-rich residual mineral phases (i.e., rutile, titanite) in the mantle source (Foley and Wheller, 1990; Pearce and Parkinson, 1993). The high Y and Yb concentrations in these magmas (rather than primitive mantle concentrations) preclude garnet as a main residual phase. These geochemical observations are consistent with the experimental studies of Conceição and Green (2004) which revealed that high potassic and shoshonitic melts can be formed by the dehydration melting of a phlogopite-pargasite lherzolite in the subcontinental mantle lithosphere at 1–1.5 GPa and 1050–1150 °C.

The timing of Arabia–Eurasia continent–continent collision, however, has been highly controversial, ranging from the Late Cretaceous (based on Late Cretaceous obduction of ophiolitic sequences, Alavi, 1994; Berberian and King, 1981; Haynes and McQuillan, 1974; Stocklin, 1974) to Late Oligocene (e.g., Agard et al., 2011), and Miocene (e.g., Berberian and King, 1981; Verdel et al., 2011). Ghasemi and Talbot (2006) suggested an Early to Middle Eocene age of collision for the Zagros suture. Karsli et al. (2012), according to zircon ages of 42 Ma for the post-collisional Sisidagi monzonitic pluton with shoshonitic affinity, in the Eastern Pontides orogenic belt, stated that the time of formation of the Neotethys suture in NE Turkey is 57–65 Ma. Our data support an Early Eocene (~50 Ma) age for the onset of collision between Arabia and Eurasia in the Western Alborz Mountains. This age matches with the I-type, subduction-related Khorram Darreh granite, in the Soltanieh area, east of Tarom, which yielded a zircon age of 53 Ma (Hassanzadeh et al., 2008). The age of collision can be taken as the time at which subduction ceased. Consequently, a minimum delay between the time of the latest active subduction plutonism (53 Ma) and the time of post-collisional plutonism (40 Ma) of about 13 Ma is obtained for this sector of the Tethyan orogen. Thermal-anomaly resulted from widespread upwelling of asthenosphere caused partial melting of chemically enriched lithospheric mantle to produce the high-potassic melts in the Tarom magmatic segment during the Upper Eocene (Fig. 15). It seems that the crustal thinning at the Western Alborz–Azarbaijan belt had been started before ~40 Ma and after ~53 Ma.

8. Conclusions

Upper Eocene high potassic to shoshonitic granitoid complexes occur in the western part of the Alborz Mountains, NW Iran. The intrusions cover a compositional range from quartz-monzodiorite through quartz-monzonite to monzogranite. The emplacement of the Tarom complex, which is I-type and has mostly metaluminous characteristics, took place

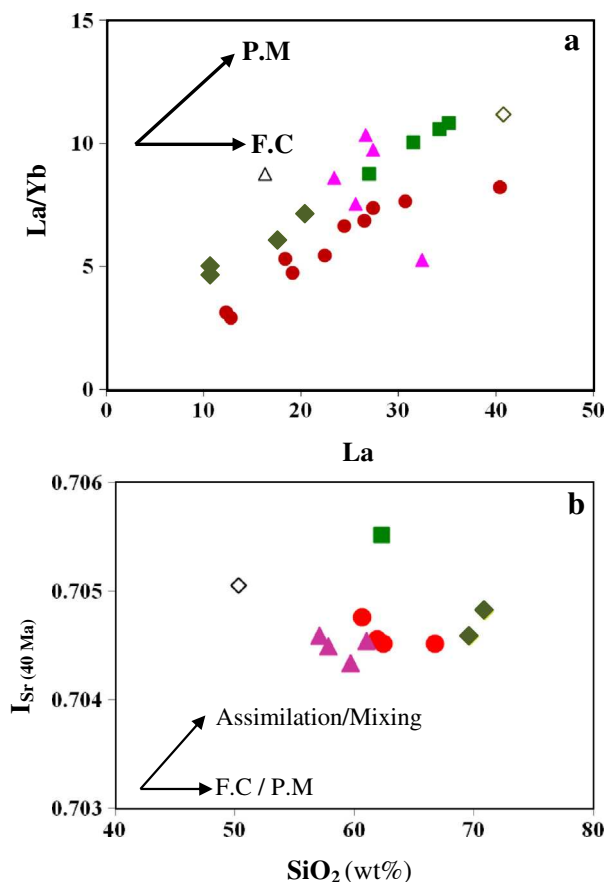


Fig. 14. Plots of (a) La/Yb ratio versus La (ppm) and (b) SiO₂ versus I_{Sr} (40 Ma) for the Tarom plutonic samples.

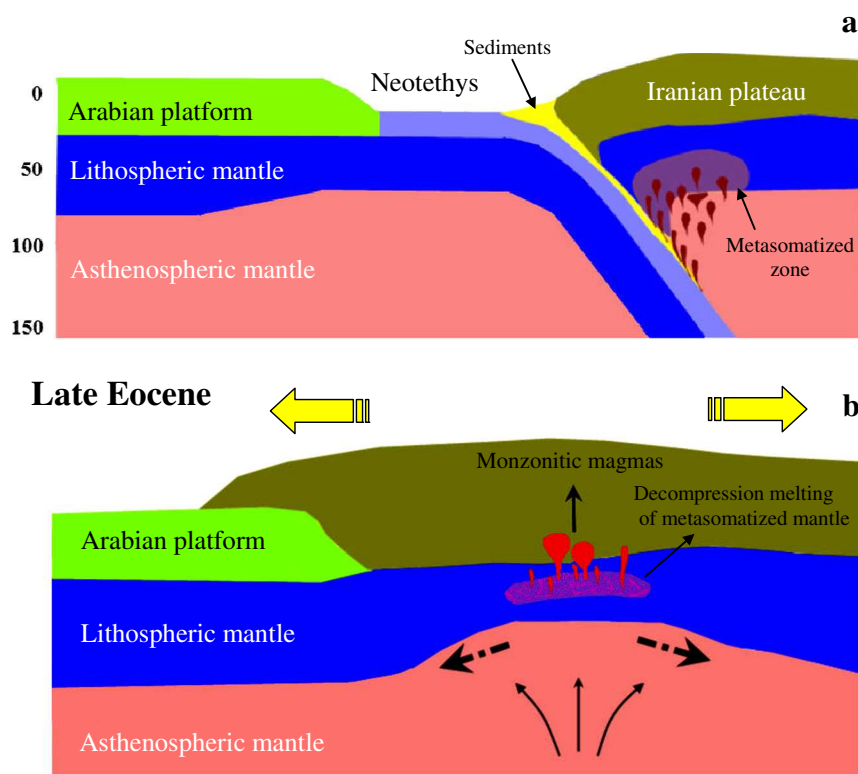


Fig. 15. Schematic illustration for the geodynamic evolution of the Tarom area: (a) formation of metasomatized and enriched zone in the subcontinental lithosphere, (b) after the collision of the Arabian–Iranian platform, the thickened lithosphere underwent convective thinning leading to asthenosphere upwelling and decompression melting of previously metasomatized mantle peridotite giving rise to the generation of high-potassic magmas.

at ~40 Ma (Bartonian) as revealed by zircon U–Pb ages. All the Tarom plutonic samples are K-rich, enriched in LREE and LILE, and depleted in HFSE and HREE. Their relatively low $I_{Sr} = 0.7043\text{--}0.7055$, moderate to low $\epsilon_{Nd} (40 \text{ Ma}) = -4.2$ to $+3.4$ and T_{DM} ages ranging from 0.7 to 1.3 Ga, with Pb isotopic ratios of $(^{206}\text{Pb}/^{204}\text{Pb}) = 18.52\text{--}18.86$, $(^{207}\text{Pb}/^{204}\text{Pb}) = 15.57\text{--}15.72$ and $(^{208}\text{Pb}/^{204}\text{Pb}) = 38.47\text{--}39.08$ are consistent with enriched lithospheric mantle source metasomatized by previously subducted sediments. Tarom plutonic rocks display some variety in Sr–Nd isotopic features which can be related to the heterogeneously metasomatized mantle source. Thermal-anomaly caused by Upper Eocene lithospheric thinning and subsequent upwelling of hot asthenospheric material beneath the Tarom area could have easily triggered partial melting of this heterogeneously metasomatized mantle, which produced magmas differing in their composition (quartz-monzodiorite and quartz-monzonite) depending on the involved mantle domain. Fractional crystallization of the Zaker quartz-monzonite resulted in formation of Morvarid monzogranitic melt. Our evidences argue against the subduction of an oceanic slab in the Tarom area postdating the early Eocene and support lithospheric thinning initiated before ~40 Ma and after ~53 Ma in this sector of the Western Alborz–Azarbaijan belt.

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