

Alkali metasomatism as a process for Ti–REE–Y–U–Th mineralization in the Saghand Anomaly 5, Central Iran: Insights from geochemical, mineralogical, and stable isotope data

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ARTICLE INFO

Keywords:

Alkali-metasomatism
Ti–REE–Y–U–Th mineralization
Saghand Anomaly 5
Central Iran

ABSTRACT

The Saghand Anomaly 5 is located in the Bafq-Saghand metallogenic province in the Central Iranian geos-
tructural zone and contains Ti, REEs, Y, U, and Th mineralization. The mineralization is related to alkali-me-
tasomatism and occurs in peripheral alteration zones around the iron ore deposits. Alkali-metasomatism and
mineralization were controlled by tectonic activity along fault zones from ductile to brittle regimes of de-
formation. A metasomatic zoning is developed from outermost unaltered lithologies (subvolcanic diorite and
quartz diorite host rocks) inwards to albite-actinolite metasomatite, phlogopite-actinolite metasomatite, and
ultimately to magnetite orebodies. The last widespread hydrothermal event corresponds to an epi-
dote₂ ± chlorite ± calcite₂ ± quartz paragenesis superimposed on all previous mineral associations.

Microscopy studies reveal that ilmenite, davidite, allanite-(La), Ti–Y–U–Th-Oxide, REE-Oxide, rutile, and
titanite and lesser amounts of thorite are ore minerals of Saghand Anomaly 5. The Ti–REE–Y–U–Th miner-
alization is related to Na–Ca-metasomatism and Ca-alteration and occurs as disseminated in albite-actinolite
metasomatites and locally as veinlet-type in aplites and granites.

The geochemical investigations indicate a calc-alkaline and continental-margin arc setting for granite and
associated diorite-quartz diorite host rocks in the Saghand Anomaly 5. The REE patterns of the mineralized
metasomatites show LREE enrichment and strong Eu negative anomalies. The presence of REE-rich hydro-
thermal phases suggests a significant mobilization and deposition of REEs, especially La, Ce and Y, by high
temperature hydrothermal solutions within the Saghand Anomaly 5.

The measured temperatures range from ~748° to ~766° C and ~703° to ~737° C for Na–Ca-metasomatic
and K–Ca-metasomatic stages, respectively, using the magnetite₁-actinolite and magnetite₂-phlogopite oxygen
isotope pairs, respectively. Both δD and δ¹⁸O values of actinolite and phlogopite indicate a magmatic source of
actinolite and phlogopite forming fluids. In addition, C and O isotopic data of calcite₂, from the late veinlets,
indicate a magmatic origin of fluids in equilibrium with calcite₂ and the formation temperature of 350 to 400° C.

1. Introduction

Hydrothermal alkali-metasomatism is quite common and occurs in a
large variety of geological environments and periods, from Archean to
Cenozoic (Cuney et al., 2012). They can be spatially associated with ore
deposits, as for some IOCG deposits (Barton and Johnson, 1996) and
Kiruna-type iron deposits or exist in barren systems (Johnson and
Harlow, 1999; Boulvais et al., 2007). There has been an ongoing in-
terest in the association of REE, radioactive elements and alkali-met-
asomatism in the last decades, as noted in several publications (i.e.,
Kinnaird, 1985; Dahlkamp, 1993; Cuney et al., 2012). Although

average U and REE contents of ore bodies associated with alkali-me-
tasomatism are not as high as in other types of ore deposits, they re-
present a promising exploration target because the resources of such
deposits are relatively large, and despite their worldwide occurrence,
they represent a significant underexplored type of mineral systems
(Cuney et al., 2012). The metasomatic and hydrothermal processes that
take place with aqueous solutions are the only endogenic processes in
which the concentration of radioactive elements, and primarily ur-
anium, reach industrial concentration (Titayeva, 1994). According to
Kinnaird (1985) Na-enrichment is accompanied by concentrations of
Fe, U, Th, Zr, Nb, Sn, Zn and HREE. Also, Na-metasomatized rocks tend

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to be enriched with respect to K-metasomatized rocks in Rb, Th, Nb, La, Ce, Hf, Zr, and Y.

The Bafq-Saghand metallogenic province in the Central Iranian microcontinent, contains the oldest (Late Neoproterozoic) basement rocks of Iran and contains world class Kiruna-type Iron Oxide-Apatite ores (IOA), Fe–Mn exhalative, Zn–Pb sedimentary-exhalative (SEDEX) types, and REE–Th–U mineralization (Daliran, 2002). The iron deposits are commonly hosted by hydrothermally altered and metasomatized rocks, which are either interstratified with volcano-sedimentary sequences, or form large subvolcanic and volcanic units (Daliran, 2002) and shallow intermediate to felsic intrusions (Ramezani and Tucker, 2003). The iron ore is dominantly a Ti–V-poor massive magnetite with subordinate hematite, and is accompanied by apatite and local REE–U–Th enrichment in some places (Daliran, 2002). The origin of the iron oxide deposit in the Bafq-Saghand mining district of Central Iran has been the subject of a long-standing dispute. Some authors believe that it was formed from magma (Samani, 1988; Förster and Jafarzadeh, 1994), while others suggest metasomatic replacement of preexisting rocks (Daliran, 1999). Moore and Modabberi (2003) suggested that the separation of an iron oxide melt and the ensuing hydrothermal processes dominated by alkali-metasomatism were both involved to different degrees in the formation of Choghart and other similar deposits in Central Iran.

Alkali-metasomatism in the Bafq-Saghand mining district occurred in several localities mainly in peripheral alteration zones of iron ore deposits. An important matter in the Bafq-Saghand mining district is the occurrence of REE–Th (such as the Choghart mine) and Ti–REE–Y–U–Th (Anomaly 5 in the Saghand mining district) mineralization in peripheral alteration zones around the iron ore deposits and the correlation between the areas containing such metals and distribution of the rocks affected by alkali-metasomatism. The close spatial and temporal association between iron deposits and Early Cambrian felsic volcanic rocks suggest that mineralization and felsic magmatism were contemporaneous (Daliran, 1990; Daliran et al., 2009, 2010).

The Uranium anomalies of Saghand area within Bafq-Saghand metallogenic province were recognized by airborne surveys in early 1978. In the year 1981, the exploration team of the Atomic Energy Organization of Iran started prospecting and exploration activities in the region which resulted in discovery of 9 anomalies, namely 1 to 9. All anomalies discovered in the Saghand area are related to alkaline metasomatism (Samani, 1993). Several Fe, U, Th, and REEs occurrences and deposits were generated in metasomatic rocks. The anomaly 5 is located in the eastern part of Saghand area and its mineralization is a paragenesis of Ti, REE, Y, U and Th elements. Mineralization occurs in albite-actinolite metasomatites above phlogopitized iron orebodies, as peripheral alteration zones around the iron orebodies. In most iron ore deposits in the Bafq-Saghand area, REE and Y elements are usually concentrated in apatite (Torab and Lehman, 2007; Rahimi et al., 2016), while the host phases of REE, Y, U and Th are quite different in the Saghand Anomaly 5.

The aim of this paper is to provide a review of the geological and tectonic setting, mineralogy, geochemistry and stable isotope analysis of Saghand Anomaly 5 in Central Iran, as well as to discuss the source and evolution of the ore-forming fluid(s) and conditions of ore formation.

2. Materials and methods

This study is based on the field observations, gamma spectrometry and sampling from various ores, host rocks, and granite intrusion bodies. A detailed petrographic investigation of a large set of samples was carried out, followed by a selection of samples for different laboratory works, including electron probe microanalysis (EPMA), whole-rock lithogeochemical analysis, and O, H and C isotopic determination. All samples were studied by optical microscope to identify the mineralogical and textural relationships.

Electron probe microanalysis (EPMA) was performed at the Iranian Mineral Processing Research Center, in Iran. EPMA was performed using Cameca SX-100 electron microprobe equipped with 5 wavelength-dispersive crystal spectrometers. Measurements were performed with an accelerating voltage of 15 kV, a beam current of 20 nA, beam size of $\sim 5 \mu\text{m}$ and 15 s counting time for major elements and an accelerating voltage of 25 kV, a beam current of 20 nA, beam size of $\sim 3 \mu\text{m}$ and 30 s counting time for trace elements. Natural oxides and silicates were used as standards and for calibration. Raw data were corrected for matrix effects with the ZAF correction procedure implemented by CAMECA.

Representative samples from various rocks were analyzed by Inductively Coupled Plasma-Optical Emission Spectrometry (ICP-OES) and Inductively Coupled Plasma Mass Spectrometry (ICP-MS) at Zarazma Mineral Studies Company, in Iran.

Isotope analyses were performed on mineral separates at the University of Oregon, USA, by using a MAT 253 gas isotope ratio mass spectrometer (IRMS) in dual inlet mode. For all isotopic measurements, only minerals without visible impurities were used. The Minerals were handpicked under a binocular microscope before isotope analysis to avoid grains with impurities. Hydrogen isotope measurements were performed on 2–3 mg of the actinolite and phlogopite minerals, in continuous flow mode using the TC/EA furnace with a glassy carbon reduction column (improved after Sharp et al., 2001). Lab protocol included vacuum-drying samples and standards in a vacuum oven at 150 °C for 12 to 24 h, followed by immediate loading and purging with the carrier gas in an autosampler. Four reference standards in three analytical sessions (NBS30 biotite, $\delta\text{D} = -66\text{‰}$, Water Canyon biotite, $\delta\text{D} = -106\text{‰}$, Butte (MT) quartz diorite biotite, $\delta\text{D} = -161.8\text{‰}$, and RUH2 muscovite, $\delta\text{D} = -98.2\text{‰}$) have been used spanning a range of 95‰ and overlapping with the range of the unknown samples. Oxygen isotopic measurements were performed on 1–2 mg of actinolite, phlogopite, magnetite_{1,2}, and calcite₂ minerals using laser fluorination (Bindeman, 2008). Samples were heated with an infrared (9.6 μm , CO₂) laser in the presence of purified BrF₅ to liberate oxygen, and the samples were not pretreated overnight to prevent room-temperature reaction. The generated O₂ gas was purified with a series of cryogenic traps held at liquid nitrogen temperature, and finally a mercury diffusion pump was used to remove any remaining traces of BrF₅. Oxygen was converted to CO₂ gas by using hot graphite, the yields were measured, and then the CO₂ gas was analyzed with the IRMS in a dual inlet mode. Four garnet standards ($\delta^{18}\text{O} = 5.75\text{‰}$, Gore Mountain, NY) were analyzed together with the unknowns during each of analytical sessions. For D/H analysis, δD average precision is $\pm 2.5\text{‰}$ and 0.05 wt% for water. Carbon isotopic numbers here are irrelevant as we convert to CO₂ gas using our in house carbon rod, the $\delta^{13}\text{C}$ values are the values of the rod. For carbonate analysis by gas bench, $\delta^{13}\text{C}$ precision is $\pm 0.1\text{‰}$ and $\delta^{18}\text{O}$ is 0.1‰.

3. Geological setting

3.1. Regional geology

The Saghand area is in the western part of the Central-east Iranian microcontinent (Fig. 1a). The Central-east Iranian microcontinent is subdivided into three fault-bounded blocks, including (from east to west) the Lut, Tabas, and Yazd blocks (Fig. 1a; e.g., Alavi, 1991). Subsequently, Ramezani and Tucker (2003) defined the Kashmar–Kerman Tectonic Zone (KKTZ) as a long, arcuate and structurally complex belt between the Yazd and Tabas blocks (Fig. 1a), coinciding with the eastern portion of the Posht-e-Badam Block of Alavi (1991) but with slightly different boundaries: the Kuhbanan–Kalmard Fault to the east and the Posht-e-Badam Fault to the west. The KKTZ is a volcano-plutonic arc (Ramezani and Tucker, 2003) that hosts the Bafq-Saghand district and the Early Cambrian volcanic and plutonic rocks of Central Iran (Fig. 1b), and is considered a distinct fault bounded zone in the

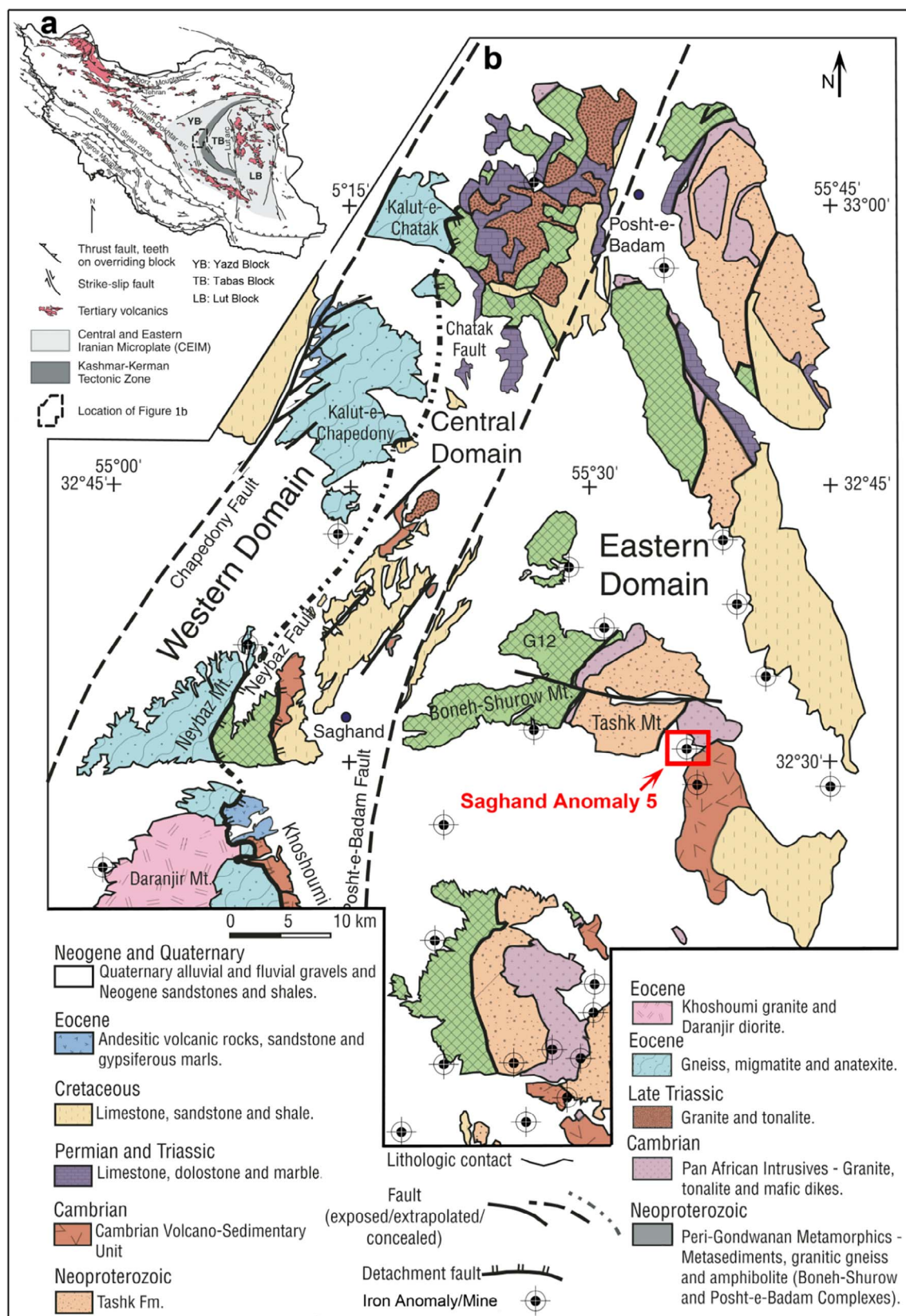


Fig. 1. (a) Location of Central and Eastern Iranian Microplate, major blocks and faults and Kashmar-Kerman Tectonic Zone (modified from Haghipour and Aghanabati, 1989), in Iran. The location of the Saghand region is outlined within the Yazd block; (b) Geological map of the Saghand area, distribution of iron deposits in the area and the location of Saghand Anomaly 5 (modified after Haghipour, 1977; Ramezani and Tucker, 2003).

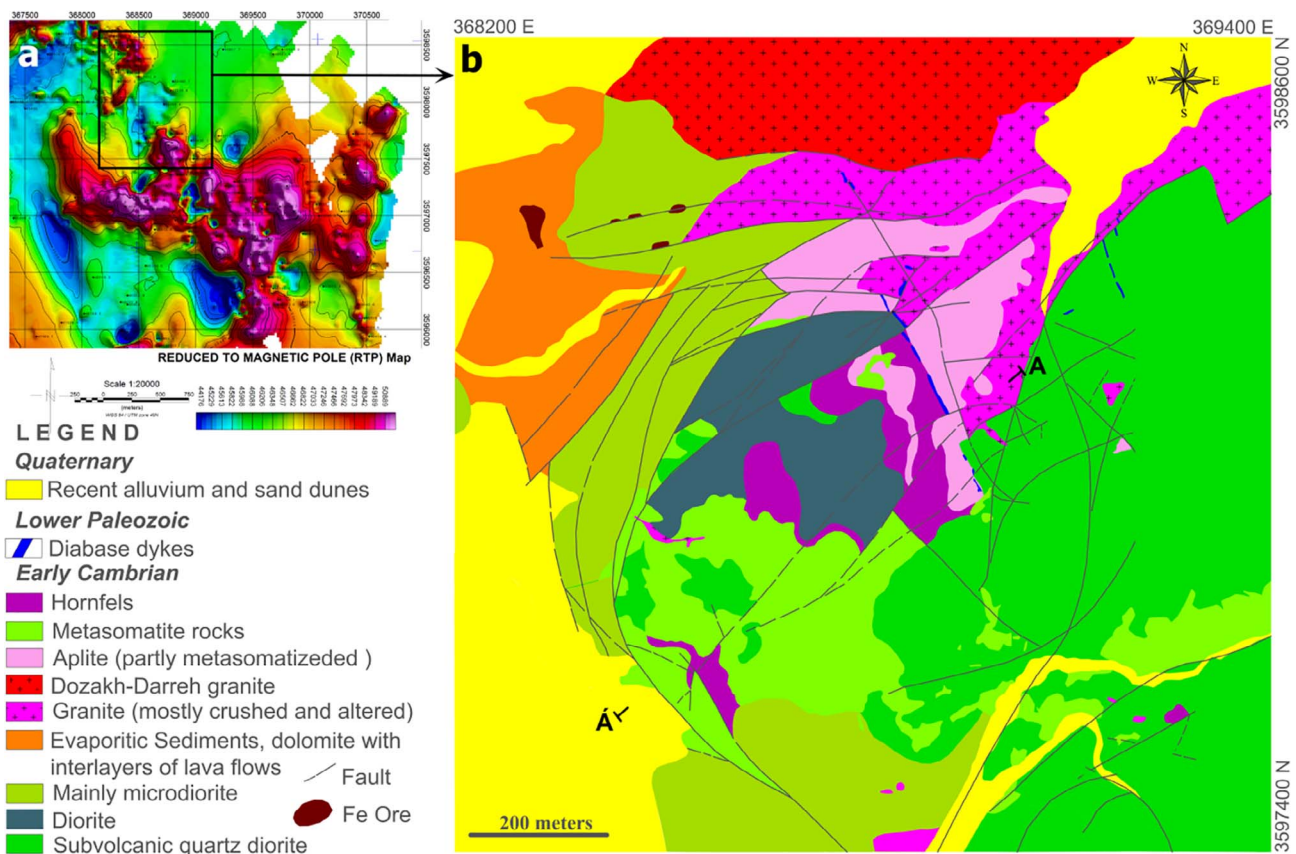


Fig. 2. (a) Reduced to magnetic pole (RTP) map of Saghand minefield (Modified after NISCO, 1979) and location of Saghand Anomaly 5; (b) geological map of the Saghand Anomaly 5 (Modified after Samani and Talezadeh Lari, 1988).

Central Iranian Terrene for several reasons (Masoodi et al., 2013) especially: (1) Unlike the neighboring Tabas and Yazd blocks, the upper Neoproterozoic and lower Paleozoic rocks are exposed along it, (2) The upper Triassic granite-tonalite plutons of the KKTZ with arc to collisional affinity are absent in the Tabas and Yazd blocks (Ramezani, 1997; Ramezani and Tucker, 2003).

The KKTZ consists of metamorphic rock units that are overlain by Mesozoic and Cenozoic units. Based on assumed stratigraphic position, Haghipour (1977) categorized these metamorphosed rocks as the Chapedony, Boneh-Shurow and Tashk metamorphic complexes from the west to east (Fig. 1). These formations are occasionally considered as one unit named Natk formation (Samani, 1993). The Chapedony Metamorphic Complex includes high-grade gneisses, migmatites, anatectic granites, marble, schist, and calc-silicate rock (Haghipour and Pelissier, 1977; Ramezani and Tucker, 2003). This complex was long considered the oldest of the Precambrian units in the Saghand region, on the basis of its high metamorphic grade (Stöcklin, 1968; Haghipour and Pelissier, 1977), but U-Pb zircon dating demonstrates a middle Eocene (46 Ma) intrusive age for the oldest components of these gneisses (Ramezani and Tucker, 2003). Further studies attributed that the complex is evolved during Tertiary to form the Eocene metamorphic core complex (Masoodi, 2006; Kargaranbafghi et al., 2007; Verdel et al., 2007; Yassaghi and Masoodi, 2011). The Boneh-Shurow Metamorphic complex, which is the most widely exposed metamorphic unit, crops out east of the Posht-e-Badam Fault (Fig. 1b). This complex is approximately 2000 m thick and exhibits a variety of metamorphic rocks, including quartz-feldspathic gneisses, green mica-schists and amphibolites (Aghanabati, 2008). Zircon U/Pb dating of this complex indicates an age of 602–617 Ma (Ramezani and Tucker, 2003). The Boneh-Shurow complex is covered by a 2000-m-thick sequence of greywackes, quartzites and quartzitic schists, marbles, amphibolites, gneiss, slaty-shales, basaltic lavas, sandstones and arkosic arenites,

argillites, tuffaceous rocks and limestones belonging to the Tashk Formation (Haghipour and Pelissier, 1977), of Late Neoproterozoic (627 Ma) to Early Cambrian age (Ramezani and Tucker, 2003) that partly metamorphosed to the greenschist facies.

The Tashk Formation is overlain by the Early Cambrian rocks in Central Iran which consists of a complex set of volcano-sedimentary sequence extends from Posht-e-Badam to Zarand and hosts the most important iron oxide $\pm$ apatite (IOA) (Fig. 1a), Zn-Pb sedimentary-exhalative (SEDEX) types, Fe-Mn exhalative and Fe-U(Th)-REE ore deposits in Iran (Samani, 1988; Rajabi et al., 2015). Ramezani and Tucker (2003) assigned an Early Cambrian age (529–554 Ma) for the volcanic suites of this volcano-sedimentary sequence and named this sequential unit as the Early Cambrian Volcano-Sedimentary Sequence (ECVSS). The volcanic rocks of ECVSS vary from felsic to mafic in composition and include rhyolite, agglomeratic tuff, splitic lava, and diabase, but dominantly rhyolitic rocks, most of which having been subject to widespread alkaline metasomatism (K-feldspar and albite; e.g., Daliran et al., 2007). The sedimentary rocks consist of dolomites, dolomitic limestones, and evaporites (Samani, 1988; Torab, 2008). The ECVSS were previously known as the Rizu, Dezu, Esfordy and Saghand Formations and were attributed to a rift-related setting (e.g., Samani, 1993, 1998; Daliran, 2002). During the Early Cambrian, granitic plutons intruded the Precambrian sequence of the Tashk Formation, and felsic to intermediate volcanic and volcano-sedimentary rocks of the ECVSS were deposited. Based mainly on their high sodium contents, Berberian and Berberian (1981) described these granitic intrusions as alkali granite. The same was considered by later researchers (e.g., Samani, 1988; Förster and Jafarzadeh, 1994; Daliran, 2002; Ghorbani, 2013); however, Torab (2008) contends that these rocks are calc-alkaline intrusions affected by widespread sodic alteration (albitization), and suggests an evaporitic source for Na. Zircon (U-Pb) dating by Ramezani (1997) indicates an age of 526 ± 3 Ma for the Zarigan,

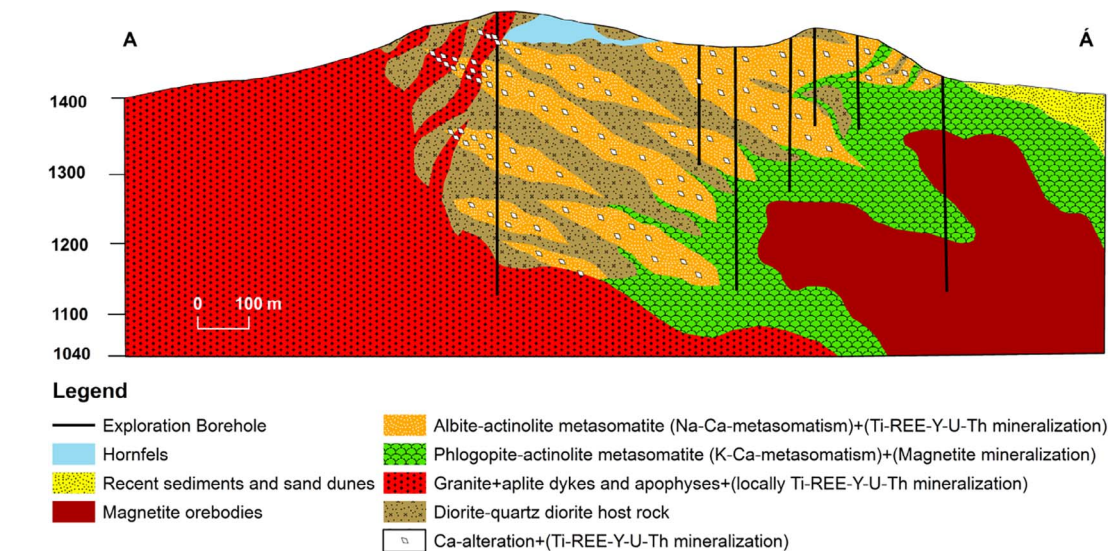


Fig. 3. Cross-section along line A–A' (in) of the Saghand Anomaly 5.

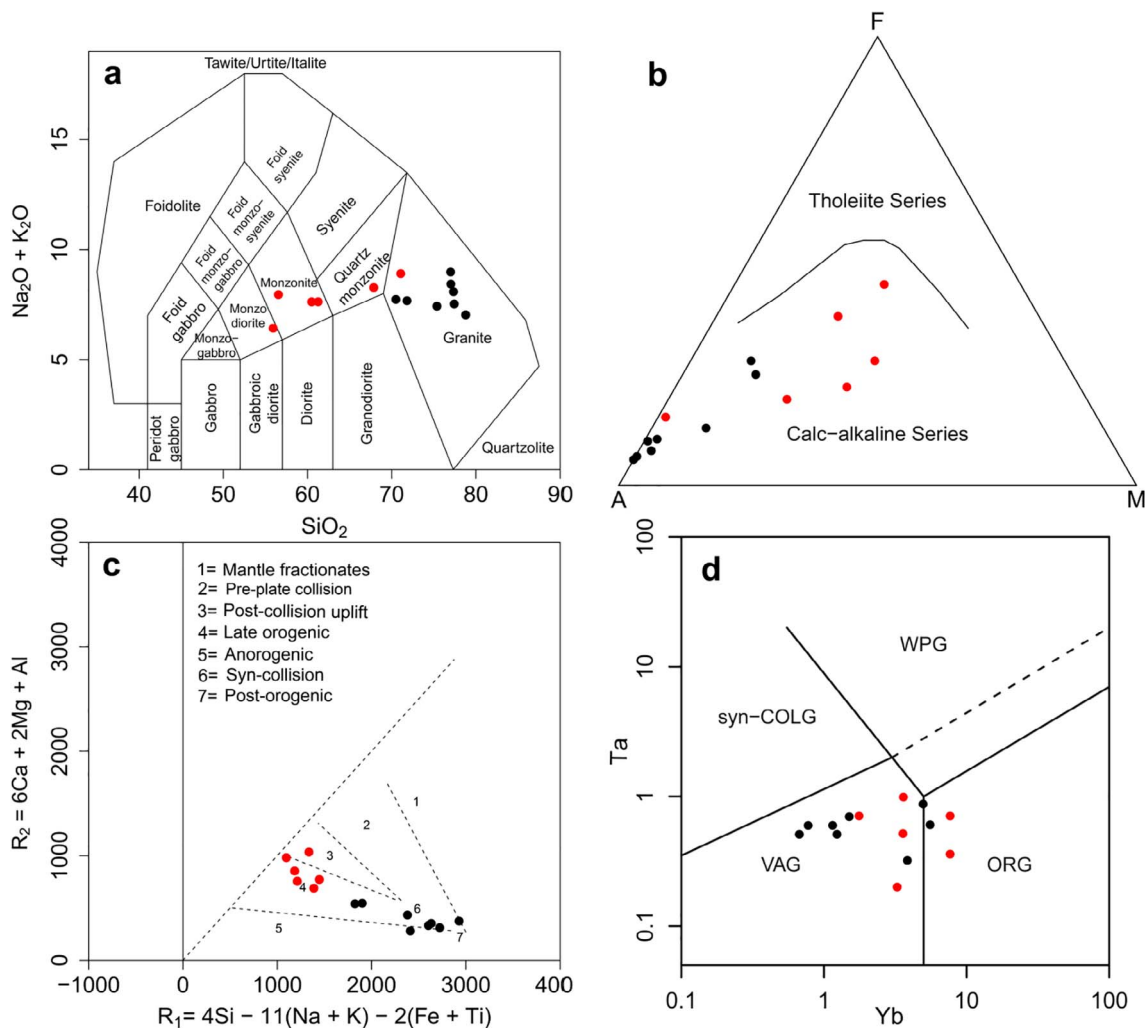


Fig. 4. (a) Na₂O + K₂O vs. SiO₂ discrimination diagram of [Middlemost \(1994\)](#) plotting samples of the igneous rocks studied; (b) Discrimination diagram for tholeiitic and calc-alkaline magma series ([Irvine and Baragar, 1971](#)) and the location of the Saghand Anomaly 5 samples; (c) R₁ versus R₂ multicatic diagram ([De La Roche et al., 1980](#)) of the geotectonic setting discrimination of granite and host rocks of the Saghand Anomaly 5; (d) The Ta-Yb discriminant diagram for syn-collision granites (syn-COLG), volcanic arc granites (VAG), within plate granites (WPG) and ocean ridge granites (ORG) ([Pearce et al., 1984](#)). The data in this study are presented in .

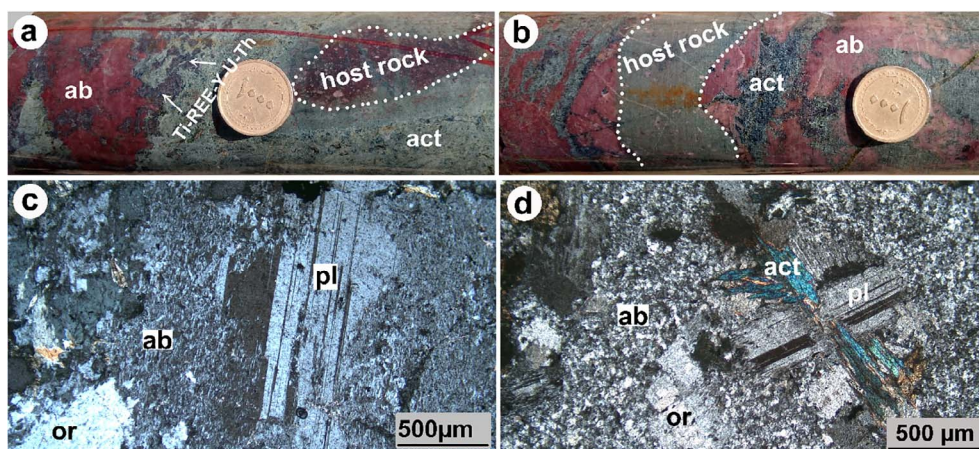


Fig. 5. Remaining evidence from (a, b) diorite and quartz diorite host rocks and (c, d) plagioclase and orthoclase minerals in metasomatized rocks of Saghand Anomaly 5 (in XPL). All abbreviations after Whitney and Evans (2010): ab = albite, or = orthoclase, pl = plagioclase, act = actinolite.

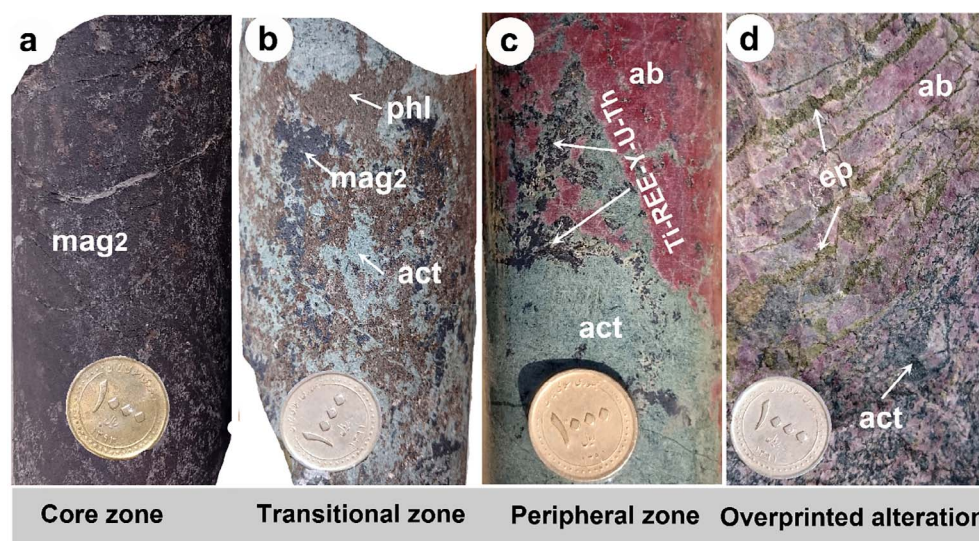


Fig. 6. Core and hand specimens photographs of metasomatic rocks representing the different zones and the evolution of metasomatic process at the Saghand Anomaly 5: (a) magnetite₂ ore in central zone; (b) phlogopite-actinolite metasomatic zone associated with magnetite₂ mineralization; (c) albite-actinolite metasomatic zone with Ti-REE-Y-U-Th mineralization; (d) epidote₂ veinlets cross-cutting the albite-actinolite metasomate representing the final stage of metasomatic evolution at the Saghand Anomaly 5. Abbreviations: mag = magnetite, phl = phlogopite, act = actinolite, ab = albite, Ti-REE-Y-U-Th = Ti, REE, Y, U and Th mineralization, ep = epidote.

Douzakh-Darreh and Sefied granites. Ramezani and Tucker (2003) ascribed these intrusions to a continental arc setting. Stosch et al. (2011) determined the U-Pb apatite age for major iron oxide-apatite deposits in the Bafq district (527–539 Ma). These ages fell entirely within the age range of the felsic plutonic rocks of the Bafq-Saghand district (including those mentioned above) dated by Ramezani and Tucker, 2003 (525–547 Ma; U-Pb on zircon).

A model proposed for the Early Cambrian tectonics and magmatism in the Bafq-Saghand district is crustal extension associated with intracontinental rifting (Berberian and King, 1981; Talbot and Alavi, 1996). It has been proposed that this rifting event formed an aborted rift associated with alkaline magmatism, including alkaline volcanic rocks and alkaline granites (e.g., Samani, 1988, 1998; Daliran, 1990, 1999, 2002; Förster and Jafarzadeh, 1994). However, advanced geochemical studies of the igneous and metamorphic rocks from the Posht-e-Badam Block by Ramezani and Tucker (2003) indicate that these rocks are attributable to a continental margin magmatic arc.

3.2. Local geology

The Saghand Anomaly 5 is mainly hosted in hydrothermally altered rocks, dominantly subvolcanic quartz diorite and diorite rocks (Fig. 2). An intrusion of medium- to fine grained leucogranite and its associated mafic (to intermediate) dikes has been emplaced into the quartz diorite and diorite rocks (Fig. 2). Many episodes of the intrusive granite emplaced into the host rocks as dike-like apophysis and aplite (Fig. 2), causing extensive alteration and mineralization in the area. The country

rocks in the vicinity of the Douzakh-Darreh Leucogranite have undergone extensive hydrothermal alteration of metasomatic type (Fig. 2). These rocks have been the subject of exploration for Fe, REE, U and Th ores. The shallow-level (subvolcanic) Douzakh-Darreh leucogranite is notably similar in composition and facies to the Zarigan intrusion which is located about 30 km to its south and west (Ramezani and Tucker, 2003). The age of 525.7 ± 1 Ma is the best estimate for the emplacement age of this leucogranite (Ramezani and Tucker, 2003). Age determination by the U-Pb method gave 507.3 Ma for the metasomatic event in Saghand area (Samani, 1988), which can indicate that mineralization in the survey area is related to the subvolcanic intrusions. In some places, there are hornfelsic rocks in the contact aureole of igneous intrusions. In the northwest of the area, there are a series of evaporitic sediments, dolomite and intermediate to felsic lava flows (Fig. 2), which are less exposed to the effects of contact metamorphism and metasomatism caused by intrusive bodies. Indeed, the mentioned lithologies are part of the Early Cambrian Volcano-Sedimentary Unit (Ramezani and Tucker, 2003) and they are known in the Saghand region as the Saghand Formation and Rizu series (Samani, 1993). The younger diabasic dikes are emplaced into the all these rocks (Fig. 2), which indicate the existence of mafic intrusions in depth. The faults and fractures of the area have two general trends of NW-SE and NE-SW (Fig. 2). The expansion of metasomatic zones often follows the NE-SW striking. Most faults and lineaments with NE-SW striking are thrust faults that are related to events after the metasomatism and mineralization and are caused the displacement of the metasomatized rocks. NW-SE striking strike-slip faults crosscut the thrust sheets and were

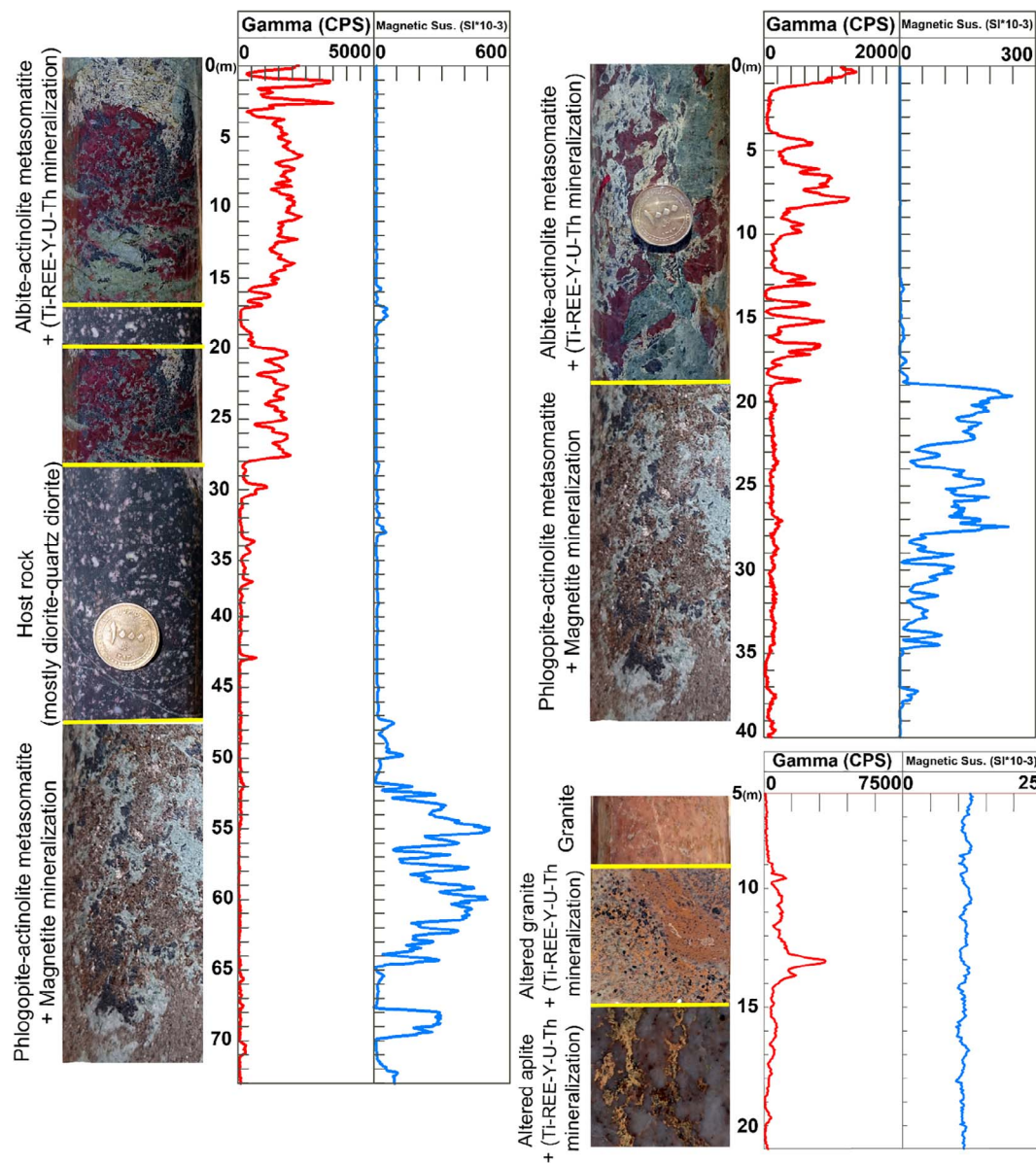


Fig. 7. Gamma spectrometry and magnetic susceptibility logs of the 3 boreholes in the Saghand Anomaly 5.

formed by right-lateral shear stresses after the mineralization.

Although the iron bodies do not outcrop at the surface of Saghand Anomaly 5, but the airborne and ground magnetometry surveys carried out in the area show large bodies of iron in depth (Fig. 2a). The drill cores in the area, especially in the southern part, also show the iron mineralization in depth (Fig. 3).

Whole rock major and trace element compositions of least-altered host rocks distal to the ore bodies were used for geochemical classification and tectonic setting. Since the subvolcanic host rocks of the Saghand Anomaly 5 were affected by hydrothermal alteration, the immobile elements, such as Al and Ti, are more suitable for petrogenetic interpretations. On the $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 discrimination diagram of Middlemost (1994), subvolcanic rocks are representative of granite and monzonite to quartz monzonite (Fig. 4a), but this is due to that these rocks were affected by extensive alteration of Na and Ca in the region and based on mineralogical characteristics, they are composed of granite, diorite and quartz diorite. A definite calc-alkaline differentiation trend (Fig. 4b) is indicated by all the samples on the $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{FeO}_t - \text{MgO}$ (AFM) ternary plot (Irvine and Baragar, 1971). On the geotectonic discrimination diagram based on

multicationic R1-R2 factor (De La Roche et al., 1980), the subvolcanic rocks plot dominantly in the late orogenic to syn-collisional fields (Fig. 4c). The Ta-Yb discriminant diagram (Pearce et al., 1984) showing volcanic arc tectonic setting for subvolcanic rocks (Fig. 4d).

4. Hydrothermal alteration and mineralization

The field relationships in Saghand Anomaly 5 are representative of a wide variety of alteration and mineralization types occurring in the host rocks (e.g. Figs. 3 and 5–7). Calc-alkaline magmatism and metasomatic phenomena progressively caused the primary rocks (mostly diorite and quartz diorite, and locally granite and aplite rocks) to change into metasomatic rocks. Ore-related metasomatic alteration drastically changed the original features, producing complex partial to pervasive replacement textures, which are increasingly altered by proximity to ore bodies. The original features, however, can be distinguished with distance from the anomaly (Fig. 2) and from the remaining evidence from primary rocks and minerals (Fig. 5).

Five main continuous stages of alteration and mineralization are identified at Saghand Anomaly 5, which exhibit zonal variations

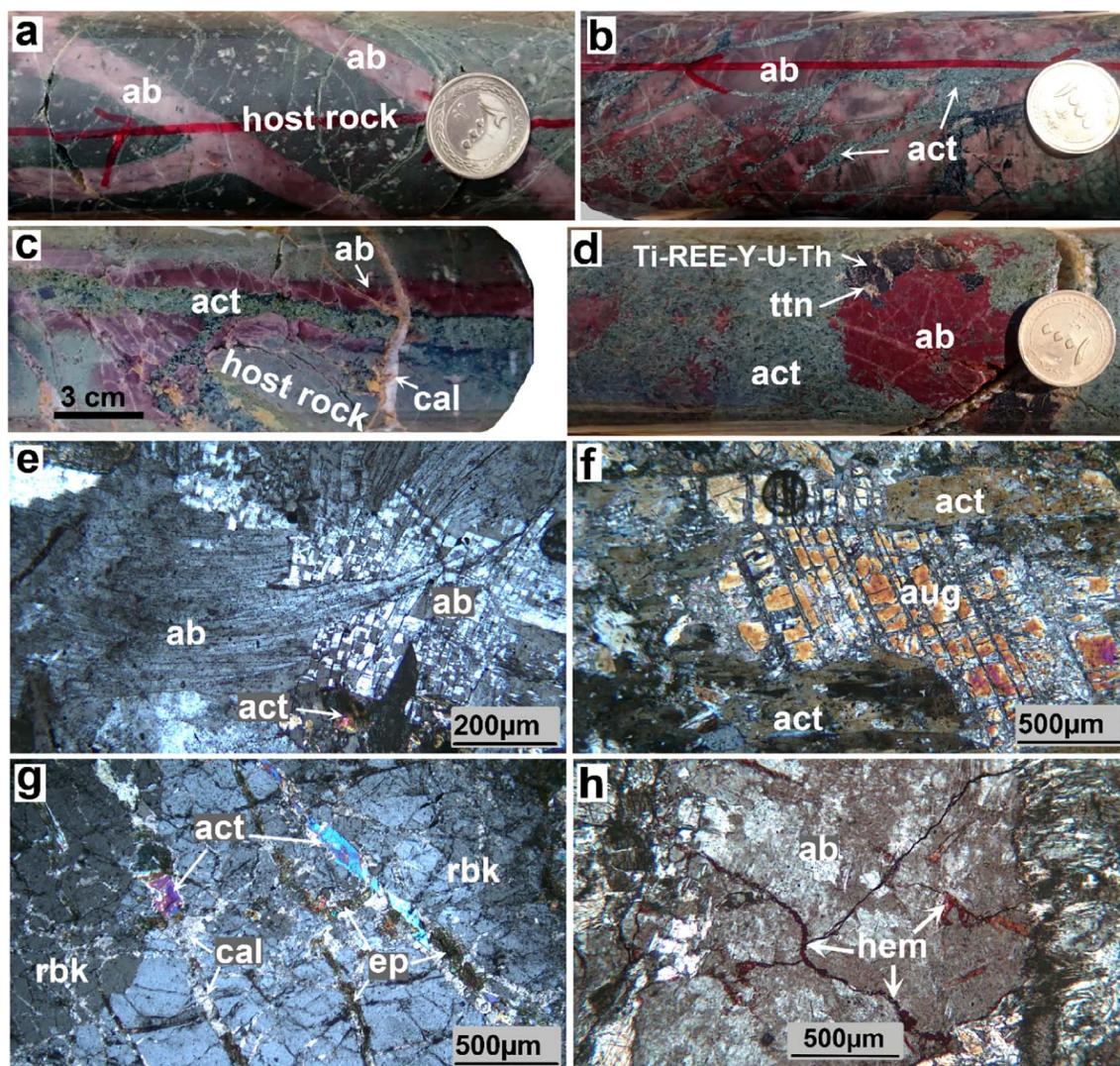


Fig. 8. (a) newly formed albite veinlets in the host rock; (b, c) brecciation of albitite and formation of actinolite veinlets; (d) patches of albitite within the actinolite metasomatite; (e) metasomatic chessboard albite and bending of albite twins in response to ductile deformation; (f) augite replaced by actinolite in latter stages; (g) cross-cutting of riebeckite by epidote₂ and calcite₂ veinlet during late alteration stage; (h) microveinlets containing fine-grained hematite inclusions in albitite. Abbreviations: ab = albite, cal = calcite, Ti-REE-Y-U-Th = Ti, REE, Y, U and Th mineralization, ttn = titanite, act = actinolite, aug = augite, rbk = riebeckite, ep = epidote, hem = hematite.

(Figs. 3 and 6). In the center of this zoning, are the iron orebodies (Figs. 3 and 6a). K–Ca-metasomatism represented by phlogopite-actinolite assemblage surrounds the iron orebodies (Figs. 3 and 6b) whereas in the outermost part of the zoning, Na–Ca-metasomatic zone is characterized by albitization and actinolitization (Figs. 3 and 6c). A Ca-alteration stage superimposed on the albite-actinolite metasomatites and comprises aggregates of Ca-minerals such as epidote₁, allanite-(La), calcite₁ and dolomite (Fig. 3). All metasomatic zones are crosscut by numerous veinlets/microveinlets of epidote₂ ± chlorite ± quartz ± calcite₂ (Fig. 6d). The intensity of metasomatism increases from the central zones towards the alteration zones on the periphery of iron orebodies. The Ti-REE-Y-U-Th mineralization is associated with Na–Ca-metasomatic and Ca-alteration stages and is hosted by albite-actinolite assemblages in the peripheral and structural zones (Figs. 3 and 6c). These zoning and lithological changes are easily identifiable and traceable in the drilled cores in Saghand Anomaly 5 (Fig. 7).

The mineralization and ore-forming process occurred along fault and fracture zones as lens-shaped bodies (Fig. 3) or as veins along fractures, or even as patches of different size at the intersections of fault and/or fractures. An individual metasomatic body of this type has a strike of N30E, and along the strike, there are several similar

metasomatic bodies, which indicate that these metasomatic bodies were pathways to fluids controlled by fault structures.

4.1. Na–Ca-metasomatism

Metasomatic activity was initiated at the Saghand Anomaly 5 by high temperature Na–Ca-metasomatism represented by albite and actinolite minerals (Fig. 8). This metasomatism type resulted in albitization and actinolitization of the subvolcanic diorite and quartz diorite host rocks and produced the albite-actinolite metasomatites. Na–Ca-metasomatism occurred in the peripheral zones and more distal to the iron orebodies especially in fault zones within the country rocks (Fig. 3). The albite and actinolite amounts vary in this zone, and sometimes the entire volume of the rock is formed of albite (albitite) or actinolite, but the metasomatized rocks are mostly formed volumetrically of actinolite. Albite is observed as vein-type, massive, spotty, and disseminated forms (Fig. 8a–d). The Na–Ca-metasomatism resulted in the formation of a fine- to coarse-grained albite, commonly displaying a chessboard texture (Fig. 8e). Albite has variably replaced Na–Ca plagioclases and K-feldspar phenocrysts and groundmass in subvolcanic rocks (Fig. 5c and d). Diopside, augite and riebeckite are minor minerals that are formed in this metasomatic stage (Fig. 8f and

Table 1

Whole rock analyses of granite, host and metasomatized rocks of the Saghand Anomaly 5.

N	ROCK TYPE	SiO ₂	Al ₂ O ₃	FeOt	MnO	MgO	CaO	Na ₂ O	K ₂ O	TiO ₂	P ₂ O ₅	Total	U	Th	V	Zr	Ta
		%	%	%	%	%	%	%	%	%	%	%	ppm	ppm	ppm	ppm	ppm
S5-212-2	Ab-act met	72.39	11.66	3.36	0.12	3.95	2.71	6.46	0.22	0.42	0.02	101.30	2018.20	1130.79	126	767	1.77
S5-215-12	Ab-act met	42.63	6.65	11.00	0.42	8.95	15.53	0.42	0.10	7.01	0.01	92.73	1763.65	1029.61	86	480	1.14
S5-503-2	Ab-act met	40.41	0.40	15.15	2.74	10.66	14.79	0.32	0.01	14.45	0.01	98.94	771.10	520.60	44	45	0.65
S5-214-13	Ab-act met	54.19	2.10	12.74	0.80	14.36	13.31	0.35	0.06	0.90	0.01	98.80	11.80	15.78	85	91	0.69
S5-214-30	Ab-act met	53.95	2.26	10.90	0.93	14.56	14.51	0.26	0.04	0.83	0.01	98.25	1.90	5.04	65	28	0.44
S5-501-2	Ab-act met	54.21	0.46	8.13	0.38	15.74	19.25	0.46	0.03	0.66	0.01	99.32	24.00	5.68	71	19	0.13
S5-213-22	Ab-act met	58.98	0.61	9.25	0.36	17.86	13.05	0.30	0.04	0.10	0.01	100.56	128.90	75.57	39	36	0.18
S5-211-23	Ab-act met	53.35	2.17	10.64	0.38	15.34	16.06	0.71	0.04	0.73	0.01	99.43	389.30	215.70	28	46	0.28
S5-503-4	Ab-act met	52.82	4.87	10.09	0.65	12.64	13.39	1.19	0.13	4.35	0.01	100.14	272.75	160.07	47	64	0.21
S5-209-5	Ab-act met	61.27	10.94	4.12	0.30	7.54	6.34	5.92	0.08	0.33	0.01	96.84	9.40	12.75	53	18	0.14
S5-212-16	Ab-act met	38.18	4.96	11.49	0.22	6.04	8.34	2.71	0.04	22.49	0.01	94.47	1.80	3.35	118	8	0.15
S5-220-2	Ab-act met	42.08	3.97	10.42	0.25	8.37	10.63	2.24	0.02	14.90	0.01	92.89	149.60	96.36	83	58	0.51
S5-209-2	Ab-act met	52.95	6.75	6.83	0.29	12.04	11.18	3.02	0.22	1.08	0.07	94.41	702.00	594.49	52	144	0.50
S5-211-17	Ab-act met	49.89	9.98	7.41	0.57	5.54	10.77	3.89	0.36	7.92	0.01	96.33	90.95	28.87	94	52	0.25
S5-211-8	Ab-act met	60.67	13.25	6.51	0.29	5.79	4.14	6.06	0.67	1.07	0.01	98.47	144.00	61.68	113	39	0.35
S5-212-12	Ab-act met	54.34	3.13	7.02	0.32	13.32	17.04	2.01	0.02	0.88	0.01	98.09	67.60	65.56	69	70	0.65
S5-212-19	Ab-act met	63.00	11.25	3.68	0.24	6.96	6.04	6.29	0.05	0.65	0.02	98.20	804.90	187.96	53	64	0.31
S5-212-29	Ab-act met	61.46	7.79	7.20	0.58	9.07	7.85	4.37	0.07	4.09	0.05	102.53	231.80	170.11	29	38	0.28
S5-212-31	Ab-act met	60.03	7.19	6.30	0.41	9.63	11.38	4.13	0.05	0.34	0.01	99.47	151.90	114.76	71	97	0.56
S5-213-7	Ab-act met	58.34	9.15	6.48	0.22	10.68	7.51	4.38	0.31	1.17	0.01	98.26	300.00	43.27	91	60	1.09
S5-214-43	Ab-act met	59.88	9.56	4.14	0.32	8.56	9.12	5.09	0.35	0.94	0.01	97.96	259.30	168.61	37	34	0.38
S5-215-10	Ab-act met	57.35	8.09	7.26	0.30	9.55	7.60	3.85	0.84	3.49	0.01	98.33	183.75	203.37	41	34	0.18
S5-215-25	Ab-act met	56.13	8.34	7.65	0.33	10.10	10.83	2.95	0.74	0.49	0.02	97.58	528.05	612.78	28	127	0.44
S5-215-26	Ab-act met	51.79	5.15	7.99	0.27	11.08	12.70	2.69	0.09	4.94	0.01	96.70	2.80	112.16	184	21	0.15
S5-221-7	Ab-act met	54.29	7.38	13.52	0.21	9.24	8.52	4.01	0.07	0.36	0.23	97.82	1.98	6.87	121	12	< 0.1
S5-504-11	Phl-act met	43.77	13.04	25.18	0.09	7.21	3.33	3.61	1.74	0.64	0.01	98.62	3.20	4.17	418	18	0.37
S5-504-3	Phl-act met	40.75	5.43	24.55	0.15	16.12	3.72	0.33	4.45	1.43	0.01	96.94	2.30	7.85	356	17	0.35
S5-213-19	Aplite	77.89	4.26	2.14	0.08	2.31	3.20	0.45	1.12	7.04	0.01	98.50	49.00	57.38	14	23	0.29
S5-213-30	Aplite	99.90	0.43	1.14	0.02	0.36	0.77	0.05	0.12	0.67	0.01	103.48	37.40	23.45	29	12	0.35
S5-214-46	Aplite	80.54	1.10	2.89	0.05	0.36	1.86	0.34	0.40	8.71	0.01	96.27	637.20	624.48	20	118	1.40
S5-210-12	Granite	75.64	10.23	0.75	0.04	0.20	1.32	6.12	0.55	0.17	0.01	95.03	3.70	26.68	22	83	0.52
S5-215-8	Granite	70.96	11.48	3.11	0.22	2.01	2.04	7.40	0.09	0.54	0.02	97.88	5.40	14.55	57	153	0.61
S5-214-34	Granite	71.04	12.94	3.63	0.16	1.77	1.83	7.54	0.16	0.65	0.02	99.74	5.26	12.12	56	147	0.86
S5-215-14	Granite	74.19	11.66	1.22	0.03	1.19	1.34	6.17	1.06	0.50	0.01	97.37	3.60	10.28	38	111	0.32
S5-210-10	Granite	74.51	11.61	0.53	0.01	0.18	0.69	2.86	4.84	0.08	0.01	95.32	2.40	22.69	14	29	0.70
S5-210-11	Granite	75.51	11.42	0.63	0.02	0.35	1.23	2.54	4.72	0.08	0.01	96.51	1.70	19.09	12	30	0.60
S5-210-7	Granite	76.35	11.80	0.50	0.01	0.17	0.85	3.78	4.48	0.11	0.01	98.06	4.10	20.04	16	46	0.61
S5-210-9	Granite	74.34	10.48	1.03	0.02	0.41	0.49	3.52	5.05	0.16	0.01	95.52	3.10	19.73	23	24	0.51
S5-504-5	Phl-act met	8.60	0.69	74.73	0.25	4.48	0.56	0.03	0.27	1.41	0.01	91.03	0.30	2.19	1396	16	0.17
S5-211-6	Diorite	55.56	13.37	11.49	0.33	7.98	1.82	5.09	1.20	1.35	0.02	98.20	1.10	3.15	171	57	0.71
S5-215-2	Diorite	55.36	15.43	6.01	0.36	8.16	2.55	4.35	3.31	1.13	0.17	96.84	2.10	5.77	131	49	0.52
S5-212-30	Quartz diorite	59.49	15.00	3.72	0.50	5.97	4.14	5.68	1.71	1.20	0.08	97.49	14.80	5.81	141	75	0.36
S5-213-12	Quartz diorite	59.98	13.16	7.42	0.33	5.06	2.29	6.17	1.19	1.25	0.07	96.92	6.16	2.93	119	22	0.20
S5-212-24	Quartz diorite	66.96	11.32	1.53	0.22	0.33	4.20	8.27	0.02	0.46	0.01	93.33	5.82	22.07	44	63	0.99
S5-213-10	Quartz diorite	66.96	12.58	2.70	0.23	3.57	3.25	8.00	0.06	0.33	0.01	97.68	9.90	10.61	81	84	0.71

Sc	Y	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Er	Tm	Yb	Lu
ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm	ppm
136.2	7075.7	9687	58	591.25	1284.1	323.25	1.58	474.71	137.19	10.28	6.80	190.12	1173.8	202.06
164.9	5400.5	7308	3201	471.98	1024.2	268.39	18.82	382.66	108.23	467.48	431.73	156.51	912.7	172.94
88.2	2680.7	3007	111	253.27	733.0	206.70	2.15	226.30	61.46	34.86	22.86	69.06	386.1	72.82
9.2	63.5	22	624	6.86	33.4	8.56	3.39	6.62	1.66	90.18	75.54	0.90	6.3	0.90
16.2	126.0	27	1233	11.91	66.1	21.17	4.28	16.44	3.91	136.99	135.29	1.59	8.4	1.20
49.9	173.1	72	416	15.14	73.7	25.95	7.06	22.29	5.57	85.78	48.03	3.12	15.2	2.48
97.6	514.4	552	29	46.26	154.6	46.42	0.69	47.31	12.82	22.17	15.84	11.67	65.6	11.89
67.2	1298.7	1531	108	123.16	338.3	92.66	2.04	104.45	27.55	34.68	19.81	31.12	185.5	33.50
91.7	844.8	1177	1755	77.79	222.6	59.57	8.13	67.88	17.90	206.24	195.17	22.57	138.7	25.32
52.8	374.2	89	83	82.43	371.1	96.83	1.38	66.16	14.84	23.38	13.06	6.09	27.3	4.44
35.4	130.3	8	10,078	4.06	27.0	10.43	25.03	10.96	3.20	1123.12	1175.92	2.23	13.1	2.66
83.3	848.9	850	7302	70.79	233.2	67.38	22.41	68.27	18.40	878.24	943.76	17.58	107.2	17.09
65.5	2460.0	3578	1033	194.93	464.6	118.09	4.61	154.30	45.53	134.30	116.52	60.19	371.2	61.23
26.6	402.3	433	3610	30.61	99.8	29.16	9.96	29.87	8.27	364.48	373.92	7.98	50.7	7.72
35.6	437.8	482	500	46.40	160.3	46.31	2.29	45.02	11.66	61.18	53.31	10.98	61.9	11.74
46.6	395.1	388	589	37.99	138.7	38.85	3.46	35.06	9.00	81.18	69.47	7.73	47.4	7.27
53.2	862.6	1762	492	116.25	369.1	93.21	2.50	86.15	20.23	62.01	51.48	23.00	149.3	26.37
90.8	884.8	1082	1769	78.07	231.6	63.07	8.61	70.90	18.83	144.46	139.78	20.88	123.6	21.57
48.3	726.6	870	1223	63.69	181.7	49.36	4.74	55.43	15.47	141.84	131.60	16.86	100.4	16.86
36.3	467.4	479	978	40.52	135.7	40.79	3.79	41.17	10.93	116.13	107.29	11.23	66.1	10.71
90.1	1025.4	1128	547	122.24	429.7	119.17	3.25	102.39	25.93	77.89	69.73	24.33	140.7	24.87
82.0	578.3	689	1409	57.18	175.3	48.27	8.19	51.34	13.16	180.99	153.73	13.23	76.4	13.89
82.8	1934.0	2647	783	177.51	424.1	109.15	4.36	143.52	39.41	96.24	84.76	49.66	289.3	51.12

(continued on next page)

Table 1 (continued)

Sc ppm	Y ppm	La ppm	Ce ppm	Pr ppm	Nd ppm	Sm ppm	Eu ppm	Gd ppm	Tb ppm	Dy ppm	Er ppm	Tm ppm	Yb ppm	Lu ppm
23.0	208.2	127	2534	47.36	199.2	43.39	8.60	28.88	5.96	312.19	306.49	2.71	17.3	2.15
57.0	181.5	30	323	18.16	97.5	31.13	2.59	23.97	5.80	35.31	20.47	2.51	13.8	1.95
1.4	11.1	25	26	1.45	6.9	1.10	0.57	1.12	0.22	1.41	1.30	0.22	3.2	0.25
5.3	48.2	12	24	4.13	25.5	7.79	0.82	5.82	1.34	8.07	5.10	0.68	5.6	0.58
10.5	544.9	123	287	46.01	244.9	115.72	9.83	88.41	27.61	179.62	112.11	15.77	77.1	11.22
4.6	166.6	372	454	42.62	164.7	41.03	2.87	31.56	6.90	42.19	28.76	4.45	26.4	4.50
55.9	2614.3	6019	7184	728.79	2386.9	533.22	34.24	438.41	87.04	569.12	467.26	75.08	445.6	76.76
1.4	9.2	9	22	1.97	12.1	2.49	0.27	1.67	0.31	1.58	1.23	0.22	1.3	0.27
12.0	57.6	32	65	7.00	33.8	8.75	1.17	6.92	1.62	10.01	6.71	0.95	5.8	0.90
12.7	54.3	19	39	3.68	21.0	6.31	0.97	5.51	1.40	8.92	5.91	0.80	5.2	0.73
9.5	34.3	16	33	3.40	21.4	6.81	0.62	4.49	1.16	6.86	3.85	0.55	4.0	0.44
1.2	17.7	17	45	5.88	30.5	7.00	0.76	3.94	0.68	3.74	1.76	0.27	1.6	0.22
< 0.5	11.9	2	5	0.28	6.8	2.13	0.30	1.61	0.37	1.91	1.18	0.17	1.2	0.19
1.2	7.4	4	8	0.39	7.1	1.92	0.43	1.19	0.26	1.24	0.74	0.13	0.8	0.16
3.1	7.3	12	28	3.25	18.6	4.43	0.80	2.28	0.34	1.58	0.88	0.13	0.7	0.15
2.6	11.3	2	< 1	0.34	6.3	1.38	0.16	1.29	0.29	1.64	1.20	0.17	6.5	0.18
20.2	10.1	10	15	0.95	8.0	1.64	0.61	1.38	0.28	1.65	1.20	0.19	1.8	0.21
29.3	34.5	30	52	4.56	23.0	6.05	1.52	4.23	0.95	5.61	3.49	0.49	3.7	0.44
30.5	70.8	62	85	6.56	31.1	9.20	1.68	7.65	1.79	11.59	8.40	1.22	8.0	1.09
22.9	31.0	23	43	4.05	20.2	4.50	0.92	3.55	0.81	4.76	3.54	0.53	3.4	0.47
7.4	36.6	22	38	4.11	23.3	6.25	0.74	4.37	1.02	5.50	3.41	0.51	3.8	0.49
8.3	102.9	37	92	12.24	60.2	16.98	1.71	13.03	3.11	18.62	10.74	1.40	7.9	1.08

Ab-act met = albite-actinolite metasomatite; Phl-act met = phlogopite-actinolite metasomatite.

g). These minerals are sometimes replaced or crosscut by actinolite, epidote₁ and calcite₁ minerals in later stages (Fig. 8f and g). Genetically, there are two types of albitites: the first in which the albitites are white to pink in color and the second that comprises the more reddish ones (meaty color). The reddish color is attributed to presence of fine-grained hematite inclusions/microveinlets in albite-rich rocks (Fig. 8h; Wilde, 2013). Field observations, geochemistry and ground spectrometry surveys, using RS-230 spectrometer, revealed that the first type has low radioactivity and mineralization, whereas the meaty ones show a high radioactivity and mineralization of Ti, REE, Y, U and Th. Presence of the meaty albite is a common aspect of albitization in the Bafq area noted in several areas (Jami, 2006; Torab, 2008; Khoshnoodi et al., 2016). According to the nomenclature of Morimoto et al. (1989) and EPMA analysis (Table 2), all pyroxenes in the Saghand Anomaly 5 belong to Ca-rich pyroxene and fall in the augite and diopside field (Fig. 9). Based on textural and microscopy evidences, such as brecciation of some albitites and their cementation by actinolite and/or crosscutting by actinolite veinlets (Fig. 8b and c), albite is sometimes formed earlier than actinolite. However, in most cases, these two minerals are paragenesis. Evidence of the granoblastic crystallization of albite (Tullis and Yund, 1985) and bending of albite twins in response to ductile deformation (Fig. 8e) indicate that Na–Ca-metasomatism was initially developed under ductile conditions.

Actinolitization has been developed as massive (Fig. 8d), vein-type (Fig. 8b, c and g), spotty and disseminated form (Fig. 6b). Fine to coarse actinolite crystals appear as needles or prismatic minerals in the hand specimens and thin sections (Fig. 10a and b). The crosscutting of albitites by actinolite veinlets (Fig. 8b and c) or brecciation of albitites and cementation by actinolite matrix, which has happened locally, indicates that Na–Ca-metasomatism was more developed under brittle conditions at the end of its period. According to nomenclature of LEAKE et al. (1997) and EPMA analysis (Table 2), the amphiboles belong to the calcic group and are actinolite (except for 1 point which falls within the tremolite field, Fig. 9b).

4.2. Ca-alteration

The Ca-alteration stage is characteristically distinguished by Ca-rich minerals such as epidote₁-allanite-(La)-calcite₁-dolomite and titanite replacing or overprinting earlier metasomatites (Figs. 8g and 10c and

d). This alteration stage is pervasive and has developed as disseminated and spotty shape in the albite-actinolite metasomatites (Fig. 10c). At this stage, ilmenite and davidite are usually altered into titanite and magnetite₃ (Fig. 10d). The Ca-alteration stage has played a very important role in the transferring and mineralization of REE, U and Th elements at Saghand Anomaly 5. The occurrence of this alteration in aplites and granites usually appears as veinlets containing dolomite, rutile, allanite-(La), Ti–Y–U–Th-Oxide and REE-Oxide. The occurring features of epidote₁ and calcite₁ in this alteration stage are different from the late alteration stage, and unlike the latter that occurs as veinlet, are disseminated and spotty, and are usually associated with mineralization.

4.3. Ti–REE–Y–U–Th mineralization

The Ti–REE–Y–U–Th mineralization is associated with Na–Ca-metasomatism (Fig. 11a and b) and Ca-alteration (Fig. 11c and d). In general, two major lithologies are the host of this type of mineralization: one as usually disseminated in albite-actinolite metasomatites (Fig. 11a and b) and another as usually veinlets in the altered aplites and granites (Fig. 11c and d). Based on field observations, spectrometric measurements, microscopy studies and geochemical results, this mineralization type does not occur during the K–Ca-metasomatic stage, and consequently, phlogopite-bearing metasomatites lack such mineralization. Therefore, this type of mineralization has a genetic relationship with Na–Ca-metasomatism. The main locality of this mineralization is the marginal fault zones containing albite-actinolite assemblages (Fig. 3).

The microscopy studies indicate that the main Ti-minerals in the Ti–REE–Y–U–Th mineralization zone are ilmenite, davidite and titanite which form euhedral to anhedral crystals in albite-actinolite metasomatite, and locally in albitite, granite and aplite (Fig. 11). Based on microscopy observations, the ilmenite and davidite minerals are formed in the Na–Ca-metasomatic stage and titanite is precipitated during the Ca-alteration stage. Ilmenite is commonly replaced by titanite and magnetite₃ from its margins and along fractures and is usually surrounded by magnetite₁ (Fig. 12). Based on electron probe microanalysis (EPMA) and optical microscopy studies, davidite ((La, Ce, Ca) (Y, U) (Ti, Fe⁺³)₂₀O₃₈) is the most abundant REE–Y–U-bearing mineral in the Saghand Anomaly 5 (Fig. 13). The total amount of REE(+Y)₂O₃ in

Table 2

Representative electron microprobe analyses of rock forming and ore minerals in metasomatites of Saghand Anomaly 5 (wt%).

Probe No	Na ₂ O	K ₂ O	MgO	CaO	MnO	FeO	Al ₂ O ₃	Cr ₂ O ₃	V ₂ O ₃	SiO ₂	TiO ₂	Y ₂ O ₃	La ₂ O ₃	Ce ₂ O ₃	Pr ₂ O ₃	Nd ₂ O ₃	Yb ₂ O ₃	Gd ₂ O ₃	Total
Augite																			
10/1.	0.34	0.02	21.64	12.85	0	5.22	0.36	0.05	0.01	57.84	0.02	0.02	0.02	0	n/a	0.06	0	n/a	98.45
22/1.	0.26	0.03	19.53	12.89	0	7.73	0.96	0.04	n/a	56.73	0.04	0.05	0.01	0.16	0	0.12	0	n/a	98.55
30/1.	0.34	0.06	19.01	12.71	0	8.07	1.1	0.06	n/a	57	0	0	0	0.09	0	0.08	0.07	n/a	98.59
31/1.	0.32	0.06	18.4	12.74	0	9.07	0.66	0	n/a	57.13	0.03	0.01	0	0.04	0.03	0.06	0	n/a	98.55
91/1.	1.03	0.11	19.45	11.86	0	8.04	0.79	0.03	n/a	57.91	0.2	0	0	0	0	0	0	n/a	99.42
26/1.	0.25	0.04	16.78	12.79	0	11.15	0.77	0	0.00	55.82	0.09	0	0	0.11	0	0.1	0.24	n/a	98.14
23/1.	0.24	0.02	20.69	13.13	0	5.05	0.23	0	0.03	56.63	0.03	0	0.04	0.16	0	0	0.07	n/a	96.32
5/1.	0.18	0.04	17.13	12.85	0	11.02	0.78	0.01	0.02	55.35	0.05	0.06	0	0	0	0.08	0.11	n/a	97.68
14/1.	0.4	0	21.48	12.88	0	5.72	0.38	0.04	0.00	57.88	0.05	0	0.1	0.02	0	0	0.05	n/a	99.00
18/1.	0.33	0.02	21.5	12.86	0	6.27	0.37	0.01	0.03	59	0.08	0	0	0	0.2	0	0	n/a	100.67
Diopside																			
13/1.	0.73	0	16.5	23.28	0	5.42	0.53	0.01	0.05	54.61	0.04	0	0	0.25	0	0.13	0.07	n/a	101.62
20/1.	0.40	0.02	14.44	25.29	0.00	4.37	0.69	0.00	n/a	54.79	0.04	0.01	0.00	0.08	0.00	0.00	0.07	0.00	100.20
102/1.	0.47	0.02	17.42	20.37	0	5.88	0.71	0.03	n/a	55.25	0.02	0.01	0.02	0.07	0	0	0	n/a	100.27
103/1.	0.42	0	17.49	23.4	0	5.17	0.16	0.04	n/a	53.06	0.08	0	0.1	0.02	0.16	0.04	0	n/a	100.14
108/1.	0.83	0	15.96	22.27	0	6.17	0.31	0.01	n/a	54.39	0.02	0	0	0.05	0	0.02	0	n/a	100.03
109/1.	0.7	0	15.92	23.39	0	5.05	0.49	0.1	n/a	54.11	0.15	0.08	0.05	0	0.03	0.05	0	n/a	100.12
17/1.	0.52	0	16.85	22.93	0	4.97	0.25	0.04	n/a	54.72	0.2	0	0	0.09	0.05	0	0.18	n/a	100.80
87/1.	0.4	0	17.07	24.15	0	4.78	0.16	0.01	n/a	53.74	0.07	0	0.06	0	0	0.05	0	n/a	100.49
90/1.	0.66	0	16.89	23.49	0	4.79	0.44	0.07	n/a	53.63	0.09	0.06	0.14	0.03	0.1	0	0	n/a	100.39
96/1.	0.38	0	17.03	23.72	0	4.96	0.19	0.05	n/a	54.75	0.05	0.03	0	0	0	0	0	n/a	101.16
97/1.	0.85	0	16.32	22.75	0	5.95	0.29	0	n/a	54.64	0.04	0.08	0	0.04	0.12	0	0.01	n/a	101.09
99/1.	0.47	0.01	16.32	23.74	0	4.85	0.24	0.04	n/a	54.72	0.07	0.01	0.07	0.07	0.15	0	0.14	n/a	100.90
Actinolite-tremolite																			
1/1.	0.36	0.02	21.53	12.79	0	5.84	0.32	0.02	0.03	59.36	0	0	0.04	0	n/a	0	0.02	0.03	100.36
105/1.	0.94	0.37	22.91	12.63	0	4.41	1.36	0	n/a	57.66	0.15	0	0	0.09	0.08	0.03	0	n/a	100.63
13/1.	0.24	0.05	19.69	13.02	0	7.71	1	0.05	n/a	56.12	0.11	0	0.03	0.06	0.03	0	0	n/a	98.11
19/1.	0.32	0.03	19.95	12.98	0	7.23	0.95	0.05	n/a	56.54	0	0.01	0	0.06	0	0	0.07	n/a	98.19
28/1.	0.51	0.19	18.74	12.29	0	8.44	1.69	0.07	n/a	56.01	0.17	0.01	0.15	0	0	0.04	0.27	n/a	98.58
29/1.	0.46	0.06	19.25	12.89	0	7.88	1.63	0.02	n/a	55.77	0.01	0.08	0.07	0.04	0	0.1	0	n/a	98.26
5/1.	0.32	0.05	19.96	12.8	0	8.51	1.74	0.04	n/a	57.39	0.19	0	0.01	0	0.02	0	0	n/a	101.03
6/1.	0.3	0.03	19.54	13.08	0	7.79	0.97	0.01	n/a	56.16	0.12	0	0	0.01	0.02	0.08	0.04	n/a	98.15
93/1.	0.44	0.09	20.02	13	0	6.85	1.36	0.02	n/a	57.22	0	0	0.14	0.19	0	0.05	0.09	n/a	99.47
22/1.	0.79	0.16	19.85	12.06	0	5.95	1.64	0	0.00	56.36	0.07	0.02	0.04	0	0	0.04	0	n/a	96.98
6/1.	0.23	0	18.52	12.85	0	7.97	1.03	n/a	n/a	54.79	0	n/a	0	0	n/a	0	n/a	n/a	95.39
7/1.	0.2	0.03	17.22	12.95	0	10.07	0.6	n/a	n/a	54.41	0	n/a	0	0	n/a	0	n/a	n/a	95.48
Davidite																			
14/1.	0.00	0.06	0.06	0.37	0.00	20.91	0.21	2.55	n/a	1.29	54.49	2.46	3.19	1.15	4.73	0.00	0.58	0.57	92.62
19/1.	0.06	0.00	0.75	0.50	0.00	19.62	0.14	0.52	n/a	0.53	55.88	3.14	3.96	3.45	5.62	0.44	0.15	0.97	95.73
100/1.	0.08	0.03	0.61	0.45	0	19.73	0.18	0.51	n/a	0.51	56.58	2.77	3.32	3.02	8.41	0.01	0.2	n/a	96.41
107/1.	0	0.02	0.75	0.52	0	21.75	0.19	0.51	n/a	0.32	55.76	3.48	0	3.84	6.95	0.32	0.17	n/a	94.58
20/1.	0.01	0	0.41	0.55	0	20.56	0.07	0.9	n/a	0.31	54.66	2.57	3.72	2.84	9.21	0.17	0.26	n/a	96.24
88/1.	0.09	0.02	0.79	0.72	0	22.36	0.17	0.57	n/a	0.72	54.46	2.98	3.17	3.12	7.94	0.25	0.43	n/a	97.79
94/1.	0.17	0.01	0.63	1.55	0	19.46	0.19	0.48	n/a	1.33	55.54	3.19	3.27	3.45	8.21	0.29	0.26	n/a	98.03
21/1.	0.02	0.02	0.96	0.66	0	25.22	0.12	0.49	2.42	0.11	54.52	2.73	3.71	3.41	6.53	0.24	0.4	n/a	101.56
12/1.	0.05	0.01	1.01	0.55	0	25.58	0.17	1.02	2.21	0.14	53.73	2.94	3.9	3.39	6.58	0.36	0.33	n/a	101.97
1/1.	0.01	0	1.24	0.63	0	24.21	0.24	1.02	2.40	0	55.82	2.72	3.66	3.62	6.11	0.27	0.32	n/a	102.27
Titanite																			
1/1.	0	0.06	0.52	27.79	0	1.62	4.95	0.06	n/a	30.46	30.63	0.45	0.01	0.08	0.05	0.13	0.07	n/a	98.46
18/1.	0.04	0.07	0.06	28.07	0	3.22	2.23	0.16	n/a	29.97	33.3	0.1	0	0.09	0	0	0.01	n/a	99.36
2/1.	0	0.27	0.68	27.45	0	1.75	4.83	0.06	n/a	30.83	30.61	0.58	0	0	0	0	0.06	n/a	98.81
21/1.	0.02	0.05	0.06	27.58	0	2.89	2.89	0.09	n/a	29.93	33.73	0.27	0.01	0	0	0.06	0.1	n/a	99.63
7/1.	0.03	0	0.03	28.3	0	1.34	4.56	0.06	n/a	30.73	32.12	0.36	0	0	0	0	0.01	n/a	99.21
89/1.	0.06	0	0.05	28.67	0	2.29	3.76	0.25	n/a	31.53	32.65	0.07	0.02	0	0	0.04	0.01	n/a	99.29
95/1.	0.01	0	0.03	28.79	0	2.57	3.81	0.26	n/a	31.59	31.8	0.03	0.1	0	0	0.2	0	n/a	99.11
98/1.	0.01	0	0	28.49	0	2.6	3.48	0.24	n/a	31.55	32.94	0.18	0	0	0	0.06	0.07	n/a	99.56
2/1.	0.04	0.01	0.03	28.49	0	1.66	3.6	0.33	n/a	30.9	32.95	0.1	0.16	0	0	0.06	0.27	n/a	98.6
9/1.	0.04	0	0.02	28.49	0	2.04	4.19	0.2	n/a	30.72	32.32	0.06	0.06	0	0.07	0.03	0	n/a	98.17
Rutile																			
34/1.	0.11	0	0	0.15	0	0.31	0.1	0.08	n/a	1.28	92.47	1.2	0	0.01	0	0.75	0	n/a	100.94
37/1.	0.02	0	0	0.04	0	0.23	0.08	0.05	n/a	0.44	96.44	0.35	0.06	0	0.01	0.13	0.06	n/a	100.9
47/1.	0	0.03	0.02	0.32	0.01	0.32	0.02	0.44	n/a	0.11	97.83	0	0.06	0	0	0	0	n/a	99.21
48/1.	0.02	0.01	0.01	0.03	0.01	0.36	0	0.24	n/a	0.14	98.39	0	0.02	0	0	0.09	0.06	n/a	99.33
54/1.	0.06	0.01	0	0.34	0	0.33	0.04	0.16	n/a	0.85	95.74	0.45	0	0	0	0.09	0.14	n/a	98.14
56/1.	0	0	0	0.14	0	0.44	0.16	0.15	n/a	0.52	95.57	0.39	0.12	0	0	0	0.02	n/a	97.51
58/1.	0.01	0.01	0	0.13	0	0.28	0.04	0.14	n/a	0.6	96.62	0.24	0.11	0.05	0	0	0.02	n/a	98.25
70/1.	0	0.01	0	0.17	0	0.31	0.09	0.15	n/a	1.17	94.59	0.78	0.05	0	0	0.24	0.08	n/a	97.4

Table 2 (continued)

Probe No	Na ₂ O	K ₂ O	MgO	CaO	MnO	FeO	Al ₂ O ₃	Cr ₂ O ₃	V ₂ O ₃	SiO ₂	TiO ₂	Y ₂ O ₃	La ₂ O ₃	Ce ₂ O ₃	Pr ₂ O ₃	Nd ₂ O ₃	Yb ₂ O ₃	Gd ₂ O ₃	Total
Allanite-(La)																			
40/1.	0.15	0.02	0.49	9.21	0	14.43	13.56	0.08	n/a	30.77	0.19	0.48	4.89	9.75	10.72	5.67	0.11	n/a	100.52
42/1.	0.24	0	0.53	8.62	0	13.14	14.24	0.15	n/a	30.69	0.22	0.27	4.18	10.3	12.56	5.99	0	n/a	101.13
43/1.	0.27	0.02	0.55	8.17	0	13.34	14.4	0.02	n/a	31.8	0.26	0.45	4.4	10.2	11.02	5.71	0.14	n/a	100.75
51/1.	0.14	0.05	0.46	8.76	0	13.94	13.67	0	n/a	28.25	0.52	0.76	5.1	11.08	10.13	6.02	0.18	n/a	99.06
52/1.	0.18	0.01	0.42	9.16	0	14.29	15.07	0	n/a	32.2	0.53	0.39	4.43	10.46	6.65	5.67	0	n/a	99.46
73/1.	0.16	0.03	0.52	8.98	0	14.38	15.06	0	n/a	31.34	0.22	0.49	4.69	9.86	8.8	6.04	0	n/a	100.57
74/1.	0.14	0.01	0.46	9.65	0	14.31	15.39	0.17	n/a	31.9	0.26	0.44	4.72	8.84	8.42	5.69	0.15	n/a	100.55
15/1.	0.04	0.02	0.05	16.98	0	13.49	20.76	0.05	0.00	36.75	0.04	0.41	3.56	5.29	2.25	0.66	0.1	n/a	100.45
16/1.	0.03	0.01	0.43	11.13	0	20.92	11.06	0	0.00	32.85	0.06	0.48	5.05	12.4	3.91	1.78	0.04	n/a	100.15
17/1.	0	0	0.1	16.49	0	15.62	17.64	0.25	0.00	35.52	0.67	1.16	4.08	4.99	3.56	0.46	0.33	n/a	100.87
Ti–Y–U–Th-Oxide																			
33/1.	0.08	0	0	0.26	0	0.27	0	0.07	n/a	0.76	49.4	16.28	0.07	0.32	0.18	2.19	0.82	n/a	70.70
35/1.	0.03	0	0	0.19	0.01	0.29	0	0.03	n/a	0.54	49.65	16.77	0	0.31	0.1	1.46	0.76	n/a	70.14
38/1.	0.11	0.02	0	0.2	0	0.38	0	0.17	n/a	0.43	49.63	16.55	0.05	0.22	0.16	1.3	0.92	n/a	70.14
46/1.	0.11	0.02	0	0.14	0.02	0.68	0	0.17	n/a	0.88	48.25	15.71	0	0.31	0	1.86	0.8	n/a	68.95
49/1.	0.05	0.01	1.23	0.17	0.04	1.01	0.79	0.14	n/a	1.65	47.58	14.49	0	0.42	0	1.61	1.04	n/a	70.23
53/1.	0.06	0.24	0.67	0.21	0	0.71	0.91	0.11	n/a	3.96	47.48	13.2	0.07	0.63	0.22	2.32	0.82	n/a	71.61
64/1.	0	0	0	0.3	0	0.54	0	0.32	n/a	0.65	48.69	15.56	0	0.38	0.19	1.22	1.05	n/a	68.90
66/1.	0.01	0.01	0	0.29	0	0.58	0	0.26	n/a	0.57	48.52	14.51	0.05	0.5	0.03	1.52	0.91	n/a	67.76
67/1.	0.01	0	0	0.3	0	0.67	0	0.27	n/a	0.74	48.99	15.23	0.08	0	0.12	1.51	1.29	n/a	69.21
75/1.	0.03	0	0	0.29	0.01	0.52	0	0.16	n/a	0.58	48.73	15.04	0.07	0.35	0.02	1.93	1.58	n/a	69.31
REE-Oxide																			
1/1.	0.00	0.00	0.00	0.28	0.02	0.00	0.00	0.00	n/a	0.20	0.07	2.44	24.18	28.61	37.29	5.49	0.07	5.78	104.44
12/1.	0.00	0.01	0.00	0.21	0.01	0.14	0.00	0.00	n/a	0.14	0.59	1.83	21.93	29.46	34.89	6.71	0.00	5.59	101.51
13/1.	0.00	0.03	0.00	0.10	0.00	0.00	0.00	0.00	n/a	0.16	1.42	1.55	21.67	29.67	33.88	7.21	0.05	5.64	101.38
17/1.	0.00	0.00	0.00	0.06	0.00	0.01	0.00	0.00	n/a	0.27	0.15	1.45	22.00	33.25	35.10	7.30	0.00	5.36	104.93
18/1.	0.00	0.01	0.00	0.03	0.01	0.00	0.00	0.00	n/a	0.20	0.06	1.11	21.44	30.99	34.26	6.34	0.00	5.47	99.92
2/1.	0.00	0.00	0.00	0.25	0.01	0.00	0.00	0.00	n/a	0.16	0.17	1.37	22.15	30.82	35.01	6.82	0.08	5.64	102.48
6/1.	0.00	0.00	0.00	0.27	0.00	0.00	0.00	0.00	n/a	0.00	0.13	1.71	20.49	29.94	33.23	6.84	0.00	5.55	98.16
7/1.	0.00	0.02	0.00	0.22	0.01	0.00	0.00	0.00	n/a	0.20	0.11	1.70	22.13	31.63	35.28	7.33	0.00	5.74	104.37
8/1.	0.00	0.01	0.00	0.19	0.00	0.00	0.00	0.00	n/a	0.12	0.06	1.60	20.98	31.03	32.66	7.00	0.00	5.48	99.14
9/1.	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	n/a	0.18	0.04	1.51	20.56	31.76	32.95	7.03	0.15	5.53	99.72

n/a: not analyzed.

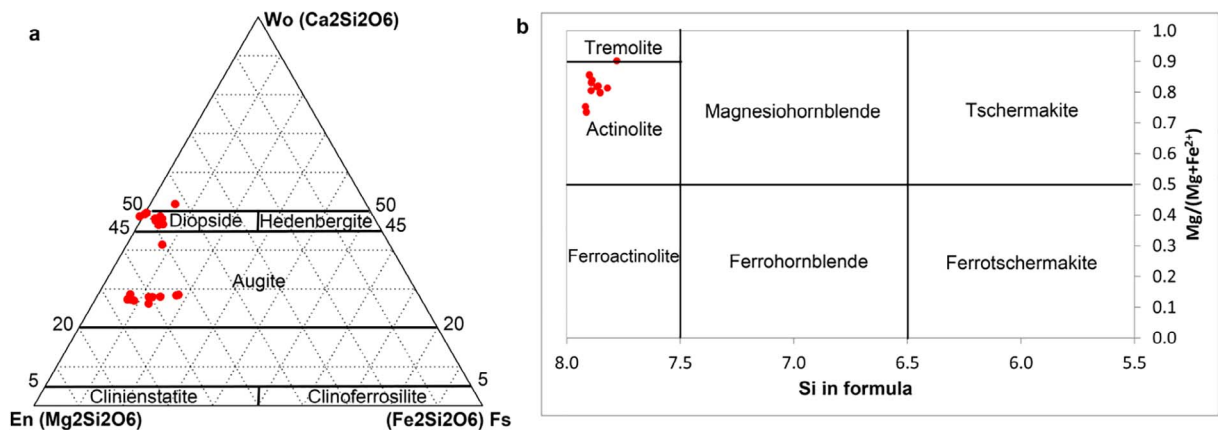


Fig. 9. (a) Classification of the Ca-rich pyroxene; (b) calcic amphiboles in the metasomatic zones of the Saghand Anomaly 5.

davidite is between 12.67 and 18.77% (Table 2). Davidite like ilmenite is generally replaced by titanite and magnetite₃ from the margin and along fractures and sometimes surrounded by magnetite₁ (Fig. 13a–e). The REE and U released due to alteration of davidite have probably contributed to the formation of allanite-(La), Ti–Y–U–Th-Oxide and REE-Oxide in the Ca-alteration stage (Figs. 14 and 15). Magnetite, in the albite-actinolite metasomatites, is commonly anhedral and is seen in two forms: the first (type 1) around and sticking to the ilmenite or davidite minerals (Figs. 12a, b, d–f, and 13a and b), which belongs to the Na–Ca-metasomatic stage, and the second (type 3) as a veinlet and scattered grains or intergrowth with titanite, which results from the alteration of ilmenite and davidite minerals in the Ca-alteration stage (Figs. 12c–f and 13c–e). Indeed, calcium- and silica-rich mineralizing

fluids cause the alteration of ilmenite and davidite that are converted to titanite. Iron released from alteration of ilmenite or davidite is precipitated as magnetite₃ within the fractures and around these minerals (Figs. 12c–f and 13c–e). Two types of titanite are recognized in the Saghand Anomaly 5: the first as euhedral titanite which occurs as individual crystals and was likely precipitated directly from the mineralizing fluids (Fig. 13f), and the second type that is anhedral and formed by alteration of ilmenite and davidite minerals (Figs. 12 and 13a–e). According to the microscopy observations, the formation of titanite is unique to the Ca-alteration stage and is derived from the alteration of the ilmenite and davidite minerals (Figs. 12 and 13). The chemical analysis results of titanite by EPMA showed that this mineral is almost devoid of rare earth elements (Table 2). Rutile is the

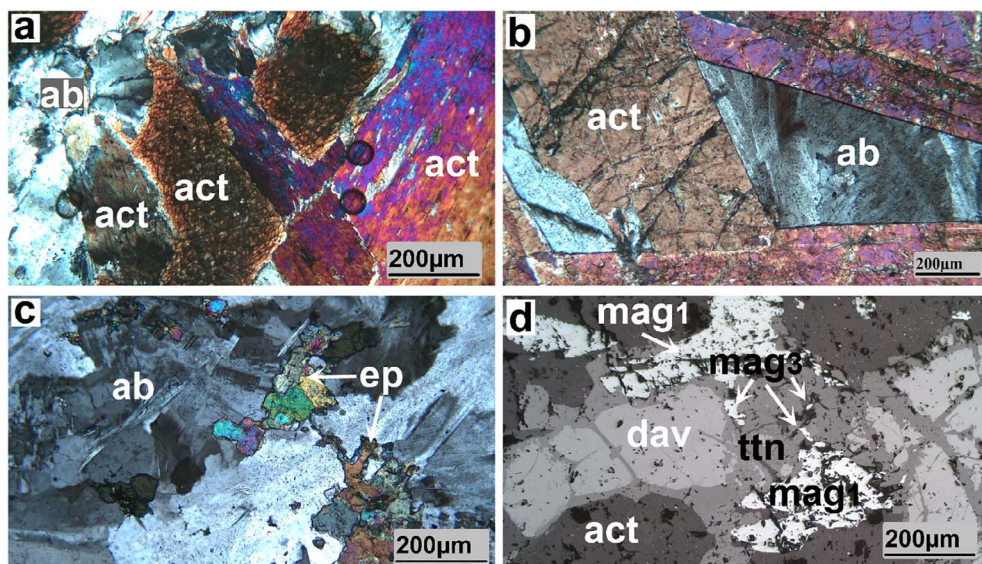


Fig. 10. Photomicrographs showing mineral paragenesis in Na-Ca-metasomatism (a, b) and Ca-alteration (c, d) stages at the Saghand Anomaly 5. (a, b) formation of actinolite prismatic crystals as paragenesis with albite in the albite-actinolite metasomatite; (c) deposition of epidote₁ as disseminated and spotty in the Ca-alteration stage; (d) davidite altered to titanite and magnetite₃ during Ca-alteration stage. Abbreviations: ab = albite, act = actinolite, ep = epidote, mag = magnetite, ttn = titanite, dav = davidite.

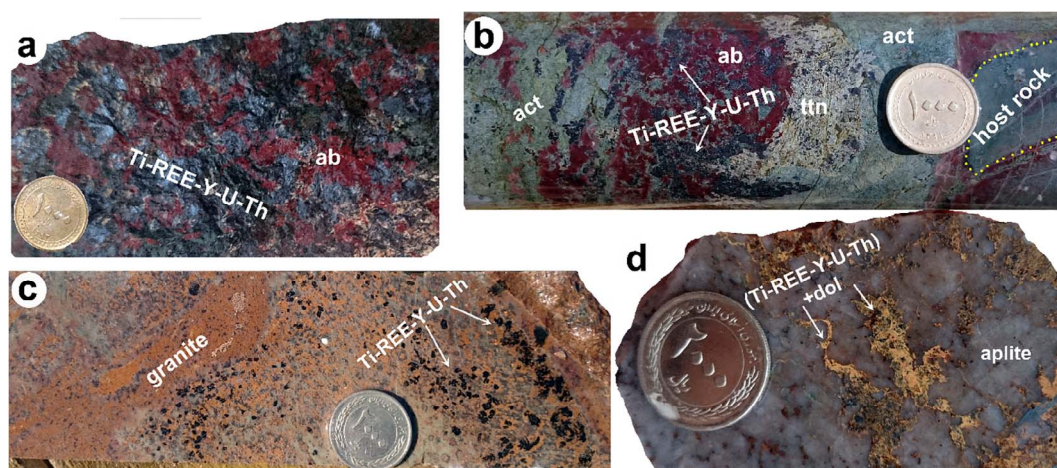


Fig. 11. The main host rocks of the Ti-REE-Y-U-Th mineralization in the Saghand Anomaly 5. (a) disseminated mineralization in albitite; (b) Ti-REE-Y-U-Th ore minerals as disseminated within albite-actinolite metasomatite and alteration of these minerals to titanite; (c, d) veinlet-type mineralization of Ti-REE-Y-U-Th ore minerals within the altered granite and aplite. These veinlets are mainly contained in rutile, Ti-REE-Y-U-Th and carbonate minerals (For more details see text). Abbreviations: ab = albite, Ti-REE-Y-U-Th = Ti, REE, Y, U and Th mineralization, dol = dolomite.

other Ti-mineral with less importance than other Ti-minerals in the Saghand Anomaly 5 that occurs commonly as veinlets in the mineralized aplites as euhedral to anhedral grains (Figs. 14 and 15). Due to the paragenesis of this mineral with dolomite, it is precipitated in the Ca-alteration stage (Figs. 14 and 15). The chemical analysis results by EPMA showed that rutile has almost no impurity in its chemical composition (Table 2).

One Ti-Y-U-Th-Oxide has been specially identified in mineralized aplite rocks (Figs. 14a, c, d, and 15a–c). It is usually anhedral and formed the intergrowth with allanite-(La), rutile and dolomite minerals (Figs. 14a, c, d, and 15a–c). The TiO₂ content of this mineral varies from 47.48 to 49.65% and its Y₂O₃ content ranges from 13.2 to 16.77% (Table 2). The U and Th contents of this mineral have not been measured (because of the lack of U and Th standards), but in the EDS spectrum, the uranium and thorium peaks are noticeable and significant (Fig. 16a and b). Another oxide mineral is also identified in mineralized aplite rocks, which is entirely composed of rare earth elements (Especially Pr, Ce, La, Nd, and Gd) and Y (Table 2). It is usually anhedral and formed the intergrowth with ilmenite, rutile, Ti-Y-U-Th-Oxide and dolomite (Fig. 14d, and 15c–f). Davidite alteration in the Ca-alteration stage has probably played a role in the formation of rutile,

Ti-Y-U-Th-Oxide and REE-Oxide (Fig. 15b, d, e). In the EDS spectrum of this mineral, also only rare earth and Y elements are noticeable (Fig. 16c and d).

Thorite (Th (SiO₄)) is commonly found as tiny and scattered minerals in albite-actinolite assemblages. This mineral is seen as inclusion minerals within the actinolite crystals under the electron microscope, which indicates the paragenesis of thorite with actinolite during the Na-Ca-metasomatic stage.

Allanite-(La) ((Ca, Ce, La, Nd, Y) (Al₂Fe³⁺) (Si₂O₇) (SiO₄) O (OH)) is the most important REE-bearing silicate mineral in the metasomatites of Saghand Anomaly 5, which is closely associated with epidote₁-calcite₁-dolomite (Figs. 15a and 17). According to the occurrence of Allanite-(La) as spotty over the albite-actinolite metasomatites (Fig. 17a–e) and its association with dolomite in mineralized aplite (Fig. 17f), it is probably formed in the Ca-alteration stage. This mineral usually has cumulative growth and is scattered as fan-shaped aggregates in the albite-actinolite metasomatites (Fig. 17a–d). The most analyzed allanite-(La) grains are heterogeneous in composition (Table 2). Allanite-(La) usually shows growth zoning (Fig. 17c and d) and the amounts of rare earth elements vary in different zones. In the back scattered electron images (Fig. 17c–e), the bright zones of allanite-

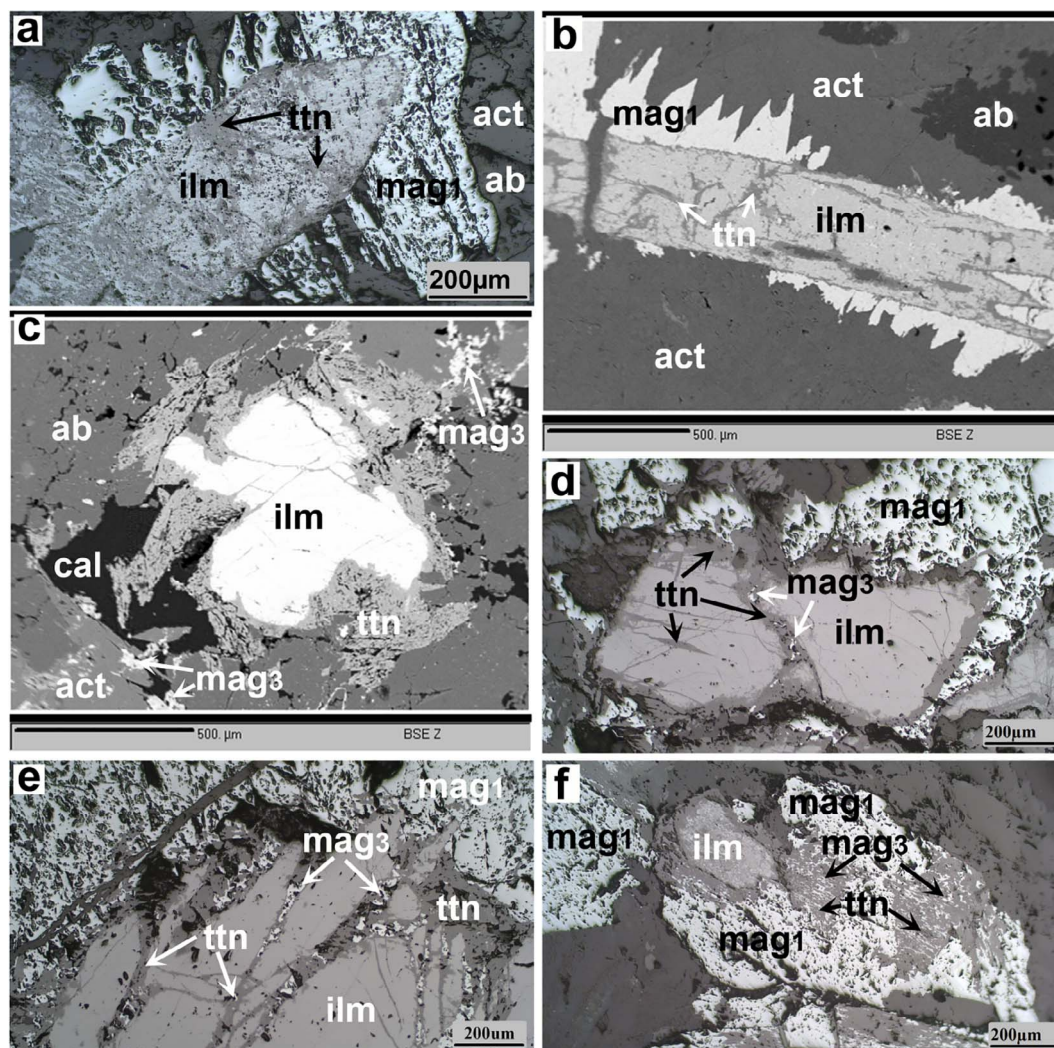


Fig. 12. Photomicrographs and EPMA-backscattered electron (EPMA-BSE) images showing the textural relationships between ilmenite, magnetite and titanite in Saghand Anomaly 5. (a, b) ilmenite crystals replaced by titanite and surrounded by magnetite; (c) ilmenite replaced by titanite and magnetite during Ca-alteration stage; (d, e, f) ilmenite minerals replaced by titanite and magnetite during Na-Ca-metasomatic stage. Note that magnetite₃ has probably derived from the ilmenite alteration in Ca-alteration stage. Abbreviations: ilm = ilmenite, ttn = titanite, mag = magnetite, act = actinolite, ab = albite, cal = calcite.

(La) have more REE content. In the aplites, allanite-(La) is also found as anhedral and fill the space between the rutile minerals (Fig. 17f).

4.4. K–Ca-metasomatism

The metasomatic alteration in the Saghand Anomaly 5 ended with the occurrence of K–Ca-metasomatism represented by phlogopite and actinolite (Figs. 6b, 8 and 18). This metasomatic stage is associated and paragenesis with iron mineralization. It overprints and replaces the Na–Ca-metasomatism (Fig. 18a). The K–Ca-metasomatism has been occasionally replaced and overprinted by quartz and other minerals in late stage of alteration (Fig. 18b and c). The K–Ca-metasomatism can be formed by alkali exchange in rocks with moderate K₂O contents (Hitzman et al., 1992). The transition from Na–Ca-metasomatism to K–Ca-metasomatism and late stage alteration can be explained due to declining temperature (Wilde et al., 2013). Based on spectrometry measurements, as well as microscopy and geochemical studies, the areas affected by this metasomatism don't have Ti–REE–Y–U–Th mineralization (Table 1) and instead, they are accompanied by magnetite₂ mineralization (Table 1, Figs. 6b, 18a, 19).

4.5. Magnetite₂ mineralization

Based on magnetometry surveys and drilled cores in the Saghand area, there are large bodies of magnetite₂ in the southern part of the Saghand Anomaly 5 (Figs. 2a and 3). Magnetite₂ orebodies locally contain major actinolite and phlogopite minerals (Figs. 6b, and 19). Based on the field observations, mineral paragenesis, and geochemical evidence, magnetite₂ mineralization at the Saghand Anomaly 5 is genetically associated with K–Ca-metasomatism. Disseminated minor sulphides, dominantly pyrite and chalcopyrite have locally occurred within the magnetite₂ orebodies. Magnetite₂ grains usually have mosaic texture and are martitized and sometimes replaced by hematite (Fig. 18d). The orebodies are essentially dominated by massive magnetite₂ (Fig. 19a), and only locally complicated by disseminated mineralization (Fig. 19a–c), and show stockwork veining and alternative banding (Fig. 19d) within K–Ca-metasomatites (Fig. 19e).

4.6. Late alteration stage

Late-stage hydrothermal vein/veinlets were developed in multiple episodes and include epidote₂ ± chlorite ± calcite₂ ± quartz, which are superimposed on and crosscut the previous metasomatic zones (Fig. 20). A breccia containing fragments of previous metasomatites,

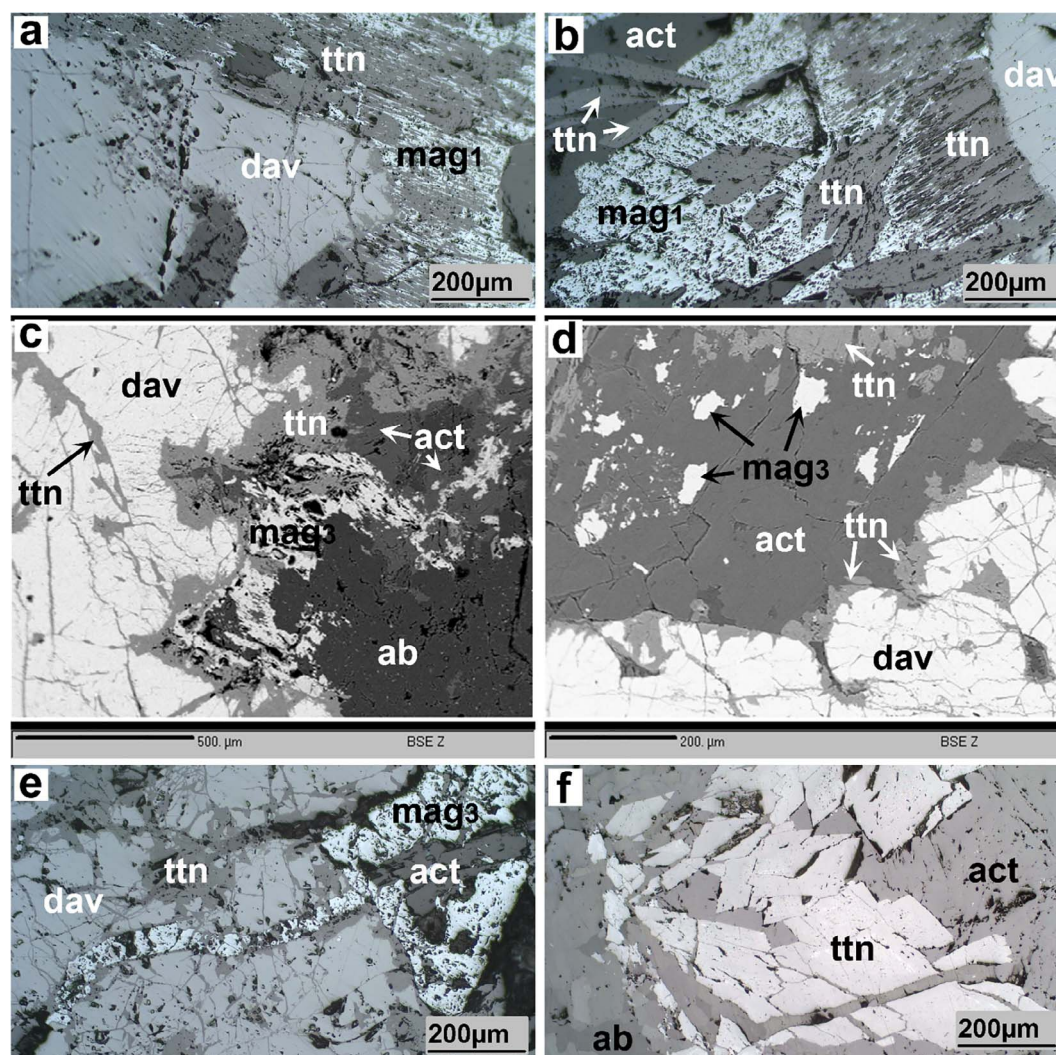


Fig. 13. Photomicrographs and EPMA-backscattered electron (EPMA-BSE) images showing the textural relationships between davidite, magnetite and titanite in Saghand Anomaly 5. (a, b) davidite replaced by titanite and surrounded by magnetite₁ during the Ca-alteration and Na–Ca-metasomatic stages, respectively; (c, d) davidite replaced by titanite and magnetite₃ during the Ca-alteration stage; (e) davidite replaced by titanite and intersected by magnetite₃ during the Ca-alteration stage; (f) euhedral titanite minerals were precipitated on the albite-actinolite metasomate during the Ca-alteration stage. Abbreviations: dav = davidite, ttn = titanite, mag = magnetite, act = actinolite, ab = albite.

and various country rocks set in a matrix dominantly of epidote₂, hematite, clay minerals and carbonates developed within some of these vein/veinlets (Fig. 20a and b). A similar breccia is reported from the Chadormalu, Esfordi and Mishdovan deposits in the Bafq district (Heidarian et al., 2017). In areas where this alteration has occurred, usually the grade of Ti–REE–Y–U–Th elements has decreased due to the leaching process by fluids. The vein/veinlets are composed of one or more minerals and abundance of epidote₂, calcite₂ and quartz present in these veins/veinlets is more than other minerals. In some of these veins/veinlets, minor amounts of specularite characterized by aggregates of flakes or tabular and anhedral crystals are found (Fig. 20f). At this stage of alteration, magnetite₂, pyrite and chalcopryrite are sometimes replaced and hematite, goethite, chrysocolla, malachite or azurite are formed instead (Fig. 20h).

Fig. 21 shows the paragenetic relationships and the paragenetic sequence of ore minerals and alteration assemblages in Saghand Anomaly 5.

5. Lithogeochemistry of metasomate, host and granite rocks

5.1. Major element variations in the Q–P mineralogical–chemical diagram

The variations in the proportions of main rock forming minerals

during the two main metasomatic stages at the Saghand Anomaly 5 can be well described in the Q–P diagram (Fig. 22). The first stage corresponds to the Na–Ca-metasomatism as albitization of K-feldspar and plagioclase as well as actinolitization, and consequently crystallization of albite and actinolite (arrow 1). The second stage corresponds to a Ca–K input with crystallization of Ca–K-rich minerals such as actinolite–phlogopite and formation of actinolite–phlogopite metasomatites (arrow 2). Introduction of Ca outside of feldspar structure leads to negative Q parameter values. Only Na–Ca-metasomatic samples have the Ti–REE–Y–U–Th mineralization (Fig. 22).

5.2. Th–U–REE geochemistry

The Th/U ratios of mineralized samples are < 1 (Fig. 23). A few metasomatized samples have Th/U values more than 1 which do not have the Ti–REE–Y–U–Th mineralization (Fig. 23). Approximately all granite and diorite-quartz diorite host rocks near the anomaly 5 have Th/U values more than 1. The low ratios of Th/U in mineralized rocks can be attributed to lower motility of Th relative to U in metasomatic fluids. Although in subduction zones, the Th content of calc-alkaline magma is usually more than U (Saunders et al., 1991), but more enrichment of uranium compared to thorium in the Saghand mineralized zone may be explained on the basis of oxidation and repeated loss of

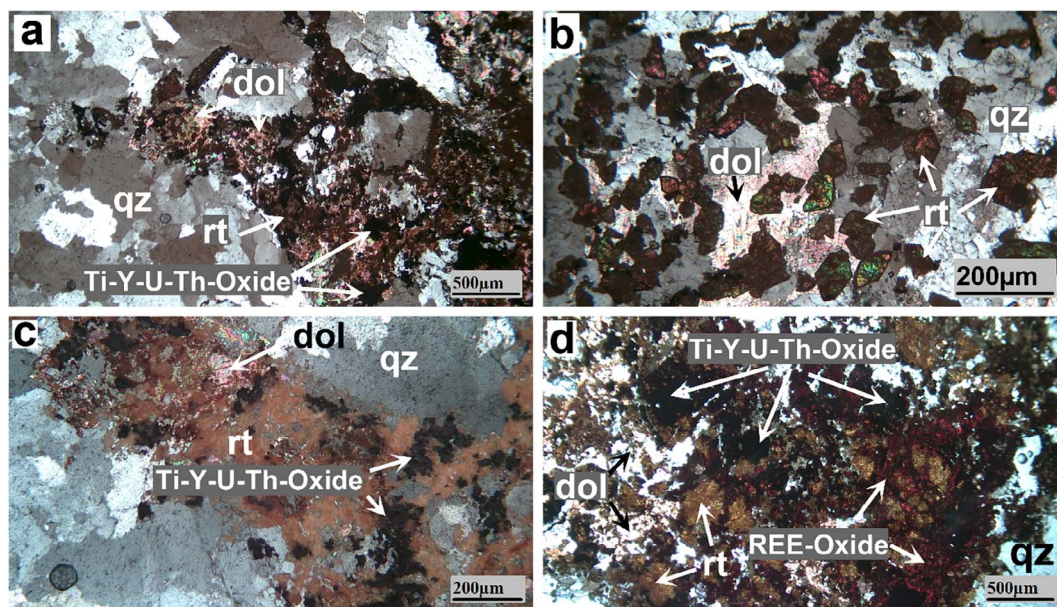


Fig. 14. Photomicrographs showing mineralization within the aplite in the Saghand Anomaly 5. (a) a veinlet containing rutile, Ti–Y–U–Th–Oxide and dolomite minerals that crosscut the aplite; (b) overgrowth of euhedral rutile, and dolomite on the aplite during the Ca-alteration stage; (c) crosscutting of aplite by rutile, Ti–Y–U–Th–Oxide and dolomite during the Ca-alteration stage; (d) paragenesis of rutile, Ti–Y–U–Th–Oxide, REE–Oxide and dolomite at the Ca-alteration stage. Abbreviations: dol = dolomite, qz = quartz, rt = rutile.

uranium from magmas during the later stages of their differentiation. If oxygen fugacity is high (greater than the Ni–NiO oxygen buffer), uranium will be greatly enriched over thorium in the fluids differentiated from magma (Bailey and Ragnarsdottir, 1994). Therefore, the oxidizing environment of hydrothermal fluid(s) in the Saghand Anomaly 5 can justify the enrichment of uranium more than thorium.

The chondrite-normalized (Nakamura, 1974) rare earth element patterns in the mineralized metasomatites exhibit notable enrichments in light REEs and heavy REEs relative to medium REEs (Fig. 24a). These rocks show negative Eu anomalies (Fig. 24a). This seems to be due to albitization of plagioclase under high oxygen fugacities (Klinkhammer et al., 1994; Cox et al., 2013). Additionally, this feature could be inherited from the fluid source as early crystallizing plagioclase in magma will remove Eu^{2+} from the system and thus the later exsolving fluid will be depleted in Eu. Hydrothermal fluids without positive Eu anomaly have been interpreted to be the result of extensive seawater contribution or mixing with detrital continental inputs (Marchig et al., 1982; Douville et al., 1999). The negative Eu anomalies in the Saghand Anomaly 5 are consistent with a contribution of an oxidized hydrothermal fluid during the mineralization process.

Saghand Anomaly 5 dioritic and quartz dioritic host rocks display slight enrichment in the LREE, with relatively flat HREE patterns and negative Eu anomalies (Fig. 24b), which reflect magmatic differentiation involving plagioclase fractionation. The REE patterns of the Douzakh-Darreh granites are very similar to those of the corresponding host rocks: a relatively flat pattern with slightly LREE enrichment relative to HREE and negative Eu anomalies (Fig. 24b). Primitive-mantle-normalized geochemical patterns (spider diagrams) of granites and diorite-quartz diorite host rocks of Saghand Anomaly 5, arranged in order of increasing element compatibility, indicate that these rocks were noticeably enriched in U and Th (large-ion-lithophile elements) and were depleted in Nb and Ta (high-field-strength elements) as well as P and Ti (Fig. 24b). The coupled negative Nb (and Ta) and positive Th anomalies in magmatic rocks, is considered a chemical signature of a subduction-zone component introduced at destructive plate boundary settings (Ramezani and Tucker, 2003). In fact, because the mantle was progressively modified by subducted oceanic crust, granites and host rocks became enriched in U, Th and LREE compared to Nb and HREE (Schandl and Gorton, 2002). Therefore, these features substantiate the origin of the granite magma as a continental-margin arc setting in the

Saghand area.

Comparison of rare earth element patterns in mineralized metasomatites with granite and diorite-quartz diorite host rocks indicates the intense enrichment of rare earth elements, especially LREEs and HREEs, in mineralized metasomatites (Fig. 24a). Therefore, during the metasomatic processes, these elements had definitely been mobile so that they were able to reach these limits of enrichment, since many studies have shown that rare earth elements are mobile during the hydrothermal and metasomatic processes (Wood et al., 1976; Taylor and Fryer, 1983; Oreskes and Einaudi, 1990; Salvi and Williams-Jones, 1990; Williams-Jones et al., 2012). For example, in the investigation of the skarn-type uranium deposits, Maas et al. (1987) deduced that REE, particularly LREE, and Th and U in the skarns, were incrementally mobile with increasing degree of metasomatic alteration. Oreskes and Einaudi (1990) in their study of Cu–U–Au–Ag–REE deposits of Olympic Dam, also reported that LREE mobilization proceeded at all the stages of alteration, while HREE mobilization was restricted to the most extreme stages of Fe metasomatism. One set of data indicates the preferential mobility of LREE (Wood et al., 1976; Campbell et al., 1984; Bence and Taylor, 1985), but there are also a number of observations, which indicate that the mobilization of HREE is dominant (McLennan and Taylor, 1979; Schneider et al., 1988). Baker (1985) found that REEs were lost in decreasing amounts with increasing atomic number during albitization.

In Saghand Anomaly 5, the enrichment of the La, Ce and Y elements is more than other rare earth elements (Table 1). Ti, U, and Th to a lesser extent also show high enrichment (Table 1). There is a positive and high correlation between these elements (Fig. 25), which indicates the association of these elements in the same mineralogical phases and their common source and genesis during the metasomatic and mineralization processes.

6. Stable isotopes

Determination of the P/T conditions of mineralization at Saghand Anomaly 5 is rendered difficult by a lack of transparent minerals amenable to fluid inclusion microthermometry, particularly in the Na–Ca and K–Ca-metasomatic stages. Based on the microscopy observations, metasomatic albite, diopside, and actinolite in the Saghand Anomaly 5 do not host inclusions of adequate size. Therefore,

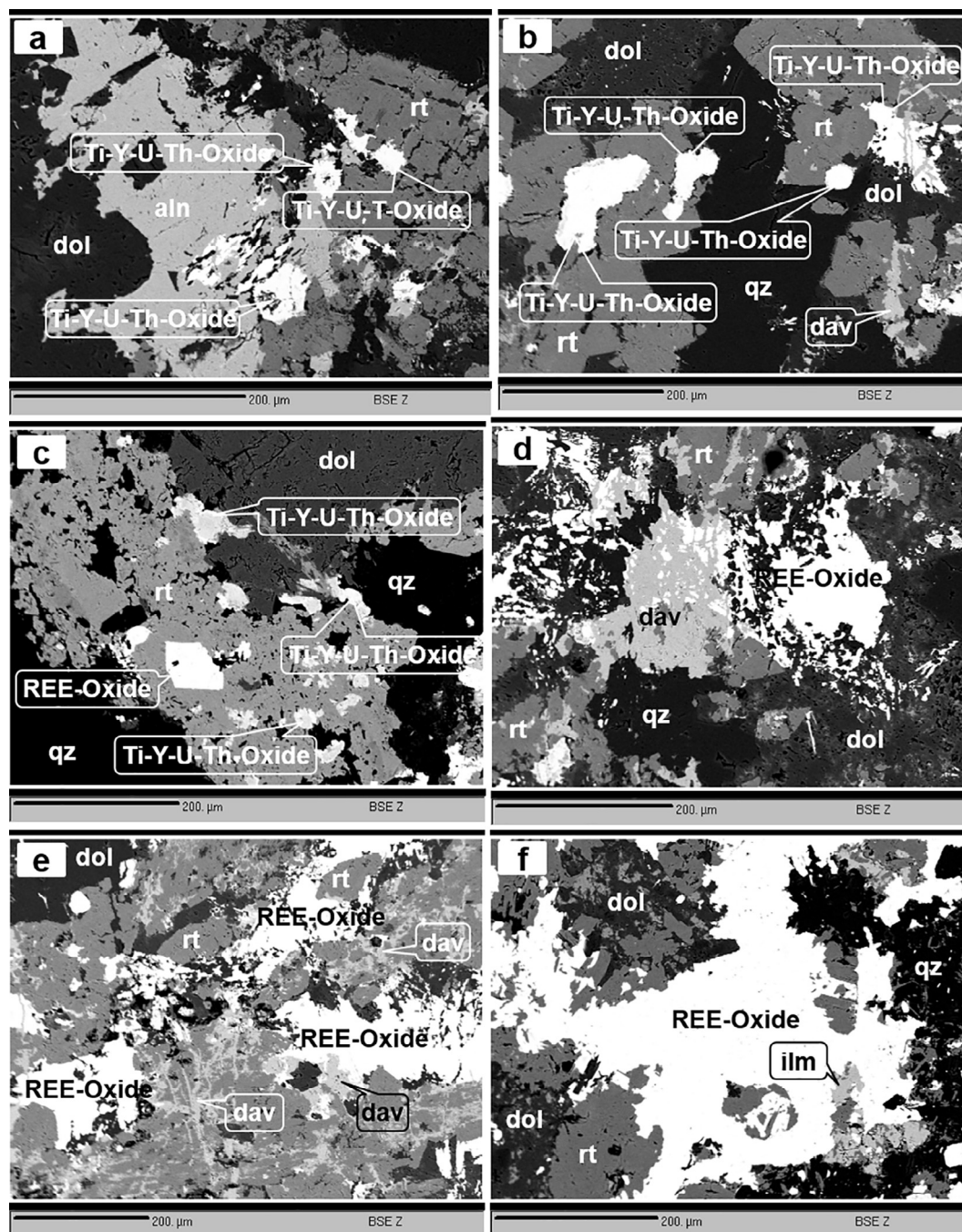


Fig. 15. EPMA-backscattered electron (EPMA-BSE) images showing mineralization in apatite at the Saghand Anomaly 5. (a, b) paragenesis of rutile, Ti-Y-U-Th-Oxide, allanite (La), and dolomite in the Ca-alteration stage; (c) formation of Ti-Y-U-Th-Oxide and REE-Oxide with dolomite and rutile as paragenesis in the Ca-alteration stage; (d, e) replacement of davidite by rutile and formation of REE-Oxide; probably REEs released from davidite alteration have contributed to the formation of REE-Oxide; (f) paragenesis of REE-Oxide, rutile, and dolomite in the Ca-alteration stage; ilmenite has been probably replaced by rutile. Abbreviations: rt = rutile, aln = allanite, dol = dolomite, dav = davidite, qz = quartz, ilm = ilmenite.

estimation of the temperatures of mineralization relies on light stable-isotope (O) fractionation between coexisting minerals. To investigate the origin and temperature of fluids involved in the various stages of metasomatism and mineralization at the Saghand Anomaly 5, some 15 samples from the drill cores were selected for oxygen, hydrogen and carbon isotopic analysis. The samples were separated by handpicking under the binocular microscope. The results are reported in Table 3 relative to standard SMOW for oxygen and hydrogen and to PDB for carbon isotopes.

As represented by mineralogy and the paragenetic assemblages, magnetite₁ can be paragenesis with actinolite in Na-Ca-metasomatism and magnetite₂ with phlogopite during K-Ca-metasomatic stage. The

MS-1 and MS-5 magnetite₂ samples and PS-2 and PS-3 phlogopite samples were derived from K-Ca-metasomatites and the AS-4 and AS-3 actinolite samples and MS-3 and MS-4 magnetite₁ samples were obtained from Na-Ca-metasomatites (Table 4). Calcite₂ minerals are taken from the veinlets which are formed at the late alteration stage. The specification of other samples is given in Table 3.

The $\delta^{18}\text{O}$ in actinolite varies from 6.01 to 7.39‰. Hydrogen isotopic compositions of actinolite in the Saghand Anomaly 5 vary from -64.4 to -70.4 ‰. The $\delta^{18}\text{O}$ isotope values vary from 6.25 to 6.72‰ for phlogopite minerals and the δD ranges from -59.3 to -66.5 ‰. Thus, both δD and $\delta^{18}\text{O}$ values of actinolite and phlogopite indicate a magmatic source for the actinolite and phlogopite forming fluids (Fig. 26a).

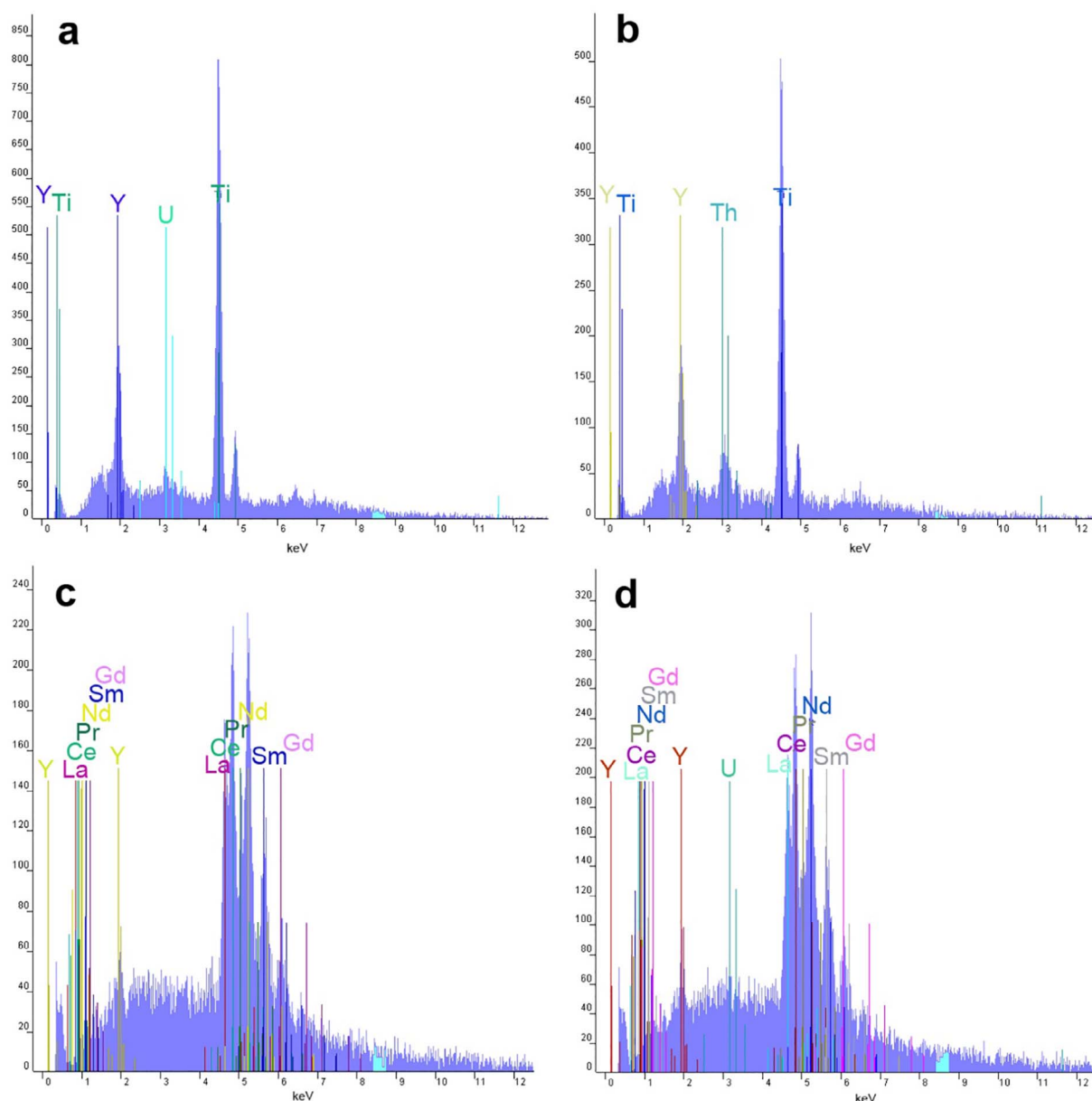


Fig. 16. (a, b) EDS spectrum of Ti-Y-U-Th-Oxide; (c, d) EDS spectrum of REE-Oxide.

The isotopic composition of hydrothermal calcite₂ varies from -5.44 to -5.53% for $\delta^{13}\text{C}$ and 10.54 to 10.64% for $\delta^{18}\text{O}$ (Table 3), corresponds to fluids with $\delta^{13}\text{C}_{\text{CO}_2}$ varying from -3.34 to -3.43% and $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ from 6.24 to 6.34% , using the experimental fractionation coefficients of Bottinga (1968) and Horita (2014) at 350°C , respectively. These values are lower than the values of marine carbonate (0% ; Faure, 1986), higher than organic material in sedimentary rocks (-30 to -10% ; Faure, 1986), slightly lower than mantle reservoir values (Ohmoto, 1972), and basically within the range of carbon in granite (Fig. 26b). Therefore, a magmatic source of carbon can be suggested for late stage of alteration in the Saghand Anomaly 5 (Fig. 26b). In addition, the $\delta^{18}\text{O}_{\text{SMOW}}$ ranging from $+10.54$ to $+10.64\%$ indicates the presence of a low temperature alteration process (Li-Qiang et al., 2017).

Isotope fractionation of oxygen between coexisting minerals, like any other stable isotope, is a strong function of temperature; for solid substances at temperatures higher than ~ 50 – 100°C , the isotope fractionation factor $1000\ln\alpha$ is a linear function of $1/T^2$ (Biegelsen and Meyer, 1947; Urey, 1947):

$$1000\ln\alpha = 10^6 A/T^2 \quad (1)$$

where A is the experimentally-determined oxygen isotope fractionation factors between mineral pairs and T is temperature in Kelvin.

When $\delta^{18}\text{O}$ values are relatively similar and not extremely low or high in $\delta^{18}\text{O}$, the fractionation factor $1000\ln\alpha$ is numerically very close to the simple difference between measured $\delta^{18}\text{O}$ values, or $\Delta^{18}\text{O}$ (e.g., Sharp, 2006; Hoefs, 2009):

$$1000\ln\alpha_{\text{actinolite-magnetite}} \approx \Delta^{18}\text{O}_{\text{actinolite-magnetite}} = \delta^{18}\text{O}_{\text{actinolite}} - \delta^{18}\text{O}_{\text{magnetite}} \quad (2)$$

and

$$1000\ln\alpha_{\text{phlogopite-magnetite}} \approx \Delta^{18}\text{O}_{\text{phlogopite-magnetite}} = \delta^{18}\text{O}_{\text{phlogopite}} - \delta^{18}\text{O}_{\text{magnetite}} \quad (3)$$

We calculated minimum formation temperatures by using the difference in O isotope composition between oxygen isotope pairs by using Eq. (1) (cf. Bindeman and Valley, 2002).

At the Saghand Anomaly 5, four oxygen isotope pairs were used to constrain the temperature of the Na-Ca-metasomatic and K-Ca-metasomatic stages (Table 4). Two actinolite-magnetite₁ isotope pairs give

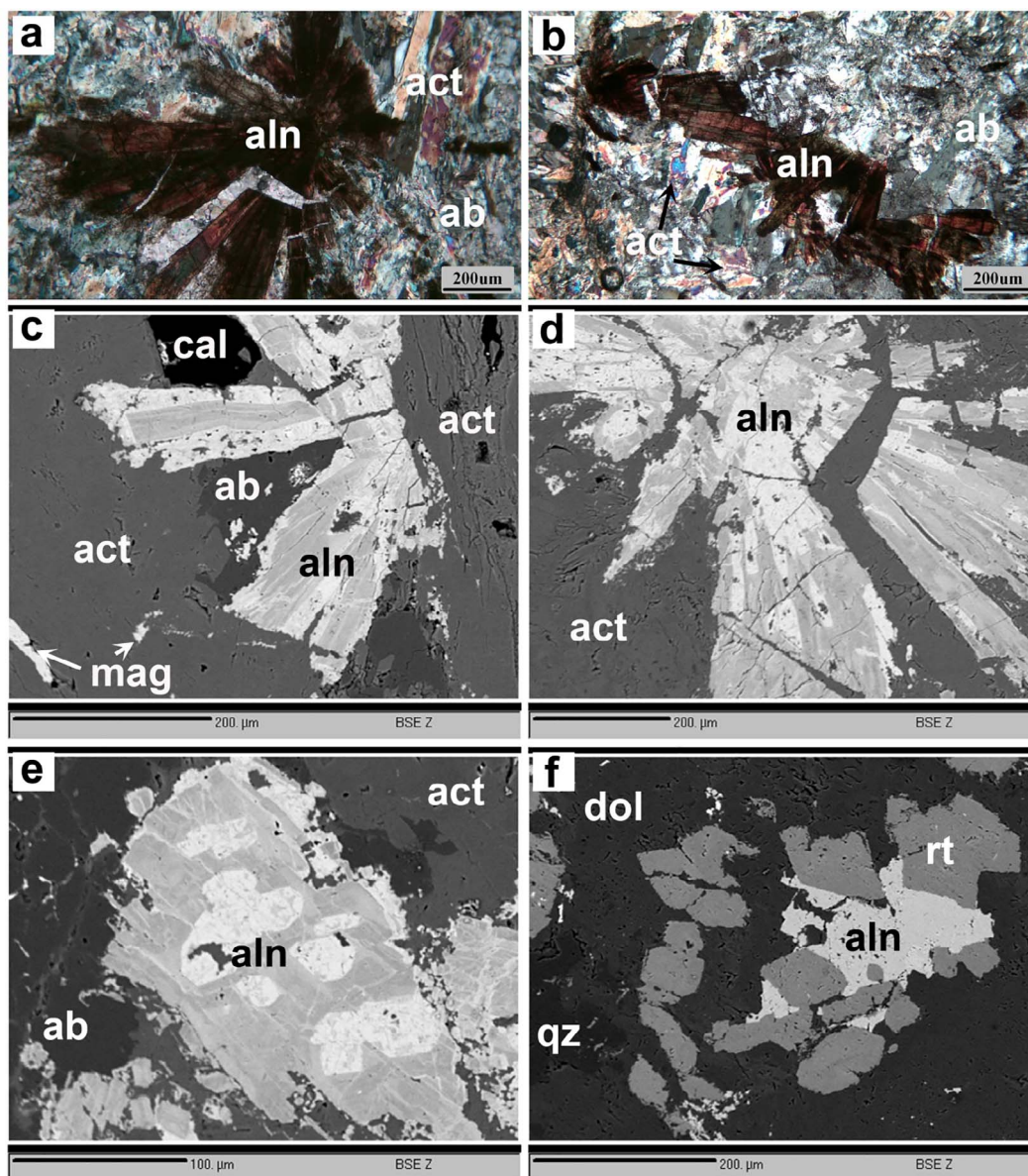


Fig. 17. Photomicrographs and EPMA-backscattered electron (EPMA-BSE) images showing the allanite-(La) mineralization at the Saghand Anomaly 5. (a, b) Fan-shaped aggregates and overgrowth of allanite-(La) minerals on albite-actinolite metasomate during the Ca-alteration stage; (c, d, e) compositional zoning in allanite-(La): the bright zones have more REE content; (f) anhedral allanite-(La) has filled the space between rutile minerals in the mineralized aplite during the Ca-alteration stage. Abbreviations: aln = allanite, act = actinolite, ab = albite, cal = calcite, mag = magnetite, dol = dolomite, qz = quartz, rt = rutile.

temperatures of $\sim 748^\circ$ and $\sim 766^\circ\text{C}$ for Na–Ca-metasomatic stage and two phlogopite-magnetite₂ oxygen isotope pairs give temperatures ranging from $\sim 703^\circ$ to $\sim 737^\circ\text{C}$ for K–Ca-metasomatic stage (Table 4), which can represent the minimum temperature of these metasomatic stages. This data implies that the hydrothermal system was cooling gradually from the Na–Ca-metasomatic stage moving to the K–Ca-metasomatic stage. These data are inferred to represent the conditions under which magnetite crystallized in equilibrium with phlogopite and actinolite based on the paragenetic relationships, although the possibility of non-equilibrium cannot be entirely excluded.

Based on the theoretical model of Zheng and Hoefs (1993), the formation temperature of calcite₂ is in the range of 350 to 400 $^\circ\text{C}$ and the water to rock ratio is about 20 (Fig. 27), which indicates that the rock isotope composition is largely controlled by hydrothermal fluid(s).

7. Discussion and conclusions

Saghand ore deposit is one of 41 iron anomalies/ore deposits

identified in Zarand-Bafq-Saghand region. The genesis of the iron deposits in the Bafq-Saghand district has long been a matter of debate. These deposits had been considered as result of liquid immiscibility, and injection or extrusion of ore magmas originated from alkaline/carbonatitic magmas, for many years (Samani, 1988; Förster and Jafarzadeh, 1994), although there is no evidence for the presence of carbonatites. Daliran (2002) concluded that hydrothermal fluids had played a significant role in the development of the Bafq-Saghand iron deposits. However, based on new investigations, it is suggested that the separation of ore rich melt and the ensuing hydrothermal processes dominated by alkali metasomatism were both involved to different degrees in the formation of ore deposits in the Bafq-Saghand district (e.g., Moore and Modabberi, 2003). Most researchers who have worked on these deposits have placed them in the category of Kiruna-type (IOA; Williams et al., 2005) Fe ore systems, and tried to provide evidence to justify the formation of these deposits via different models presented for Kiruna-type deposits (e.g., Daliran, 2002; Moore and Modabberi, 2003; Heidarian et al., 2016).

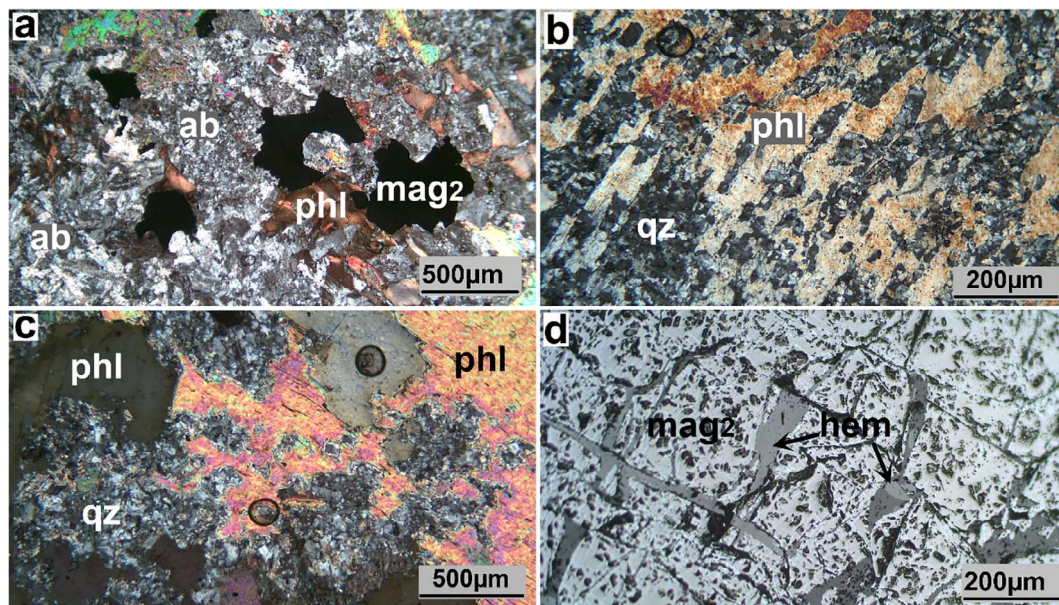


Fig. 18. Photomicrographs showing various textural relationships in the Ca-K-metasomatic zone. (a) the overgrowth of phlogopite and magnetite₂ on albite; (b, c) Phlogopite replaced and overprinted by quartz during the late stage alteration; (d) the mosaic texture of magnetite₂ grains which are martitized and replaced by hematite. Abbreviations: ab = albite, mag = magnetite, phl = phlogopite, hem = hematite.

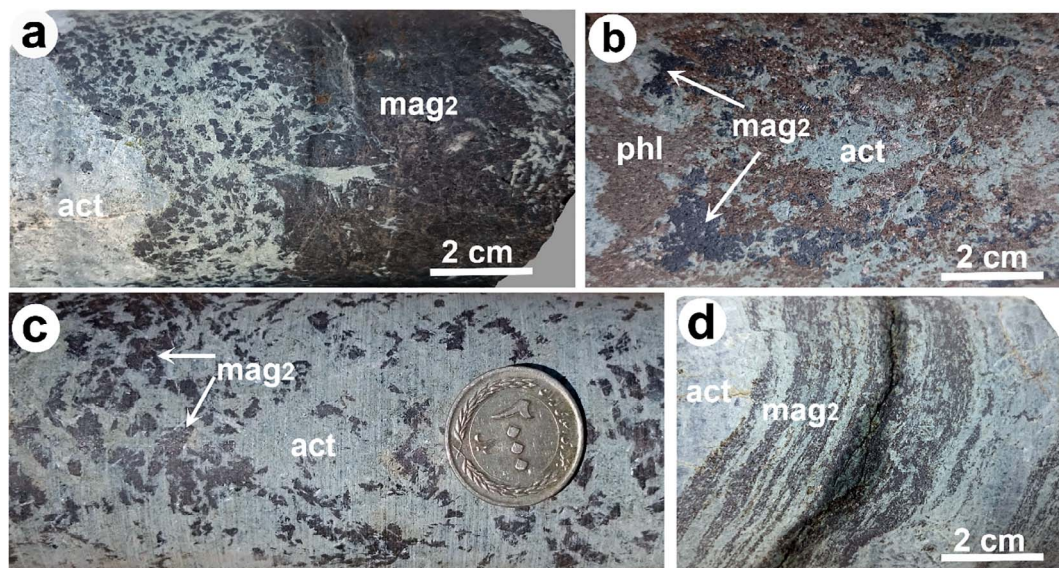


Fig. 19. The various forms of magnetite₂ occurrence during K-Ca-metasomatic stage at the Saghand Anomaly 5: (a) massive to disseminated; (b, c) disseminated; (d) alternative banding. Abbreviations: mag = magnetite, phl = phlogopite, act = actinolite.

As mentioned earlier, due to the lack of suitable minerals for the study of the fluid inclusions, stable isotopes were used to determine the source and temperature of the fluid (fluids) involved in the formation of Saghand Anomaly 5. In this study, field and mineralogical observations, as well as oxygen, hydrogen and carbon isotopic data, and other geochemical patterns have been used to construct a genetic model for the Saghand Anomaly 5 and associated metasomatic alteration.

7.1. Source, transferring and deposition of fluid(s) and metals

The origin of the Na-rich fluids, commonly accompanied by Ca-bearing minerals (actinolite, calcite, dolomite), is contentious. Albitization of rocks may be the consequence of hydrothermal fluid circulation around sea-floor hydrothermal cells (e.g., Rose et al., 1992). Alternatively, metamorphic recycling of evaporitic components is argued to be responsible for the formation of albite-rich rocks (Cook and

Ashley, 1992; Oliver et al., 1994, 2004). Barton and Johnson (1996) proposed that circulation of saline, evaporate-derived brines was instrumental in stripping iron and base metals from wall rocks, which are left behind with intensive sodic-calcic alterations. Pollard (2001) and Sillitoe (2003), however, described this alteration as a product of highly saline fluids derived from intermediate to felsic magmas. A meteoric or formation water origin for the albitizing fluids has been already proposed for some albitite occurrences in the world, as for example Polito et al. (2009) for the Valhalla uranium deposit in Australia, Johnson and Harlow (1999) for albitites in Guatemala, Lobato and Fyfe (1990) for the Lagoa Real uranium deposit in Brazil, Alexandre (2010) for the uranium mineralized albitites of Kurupung, Guyana.

The isotopic composition of oxygen, hydrogen and carbon in phlogopite, magnetite, actinolite and calcite₂ minerals at the Saghand Anomaly 5 is in the range of magmatic waters and hence provides further support for a magmatic hydrothermal origin for these minerals

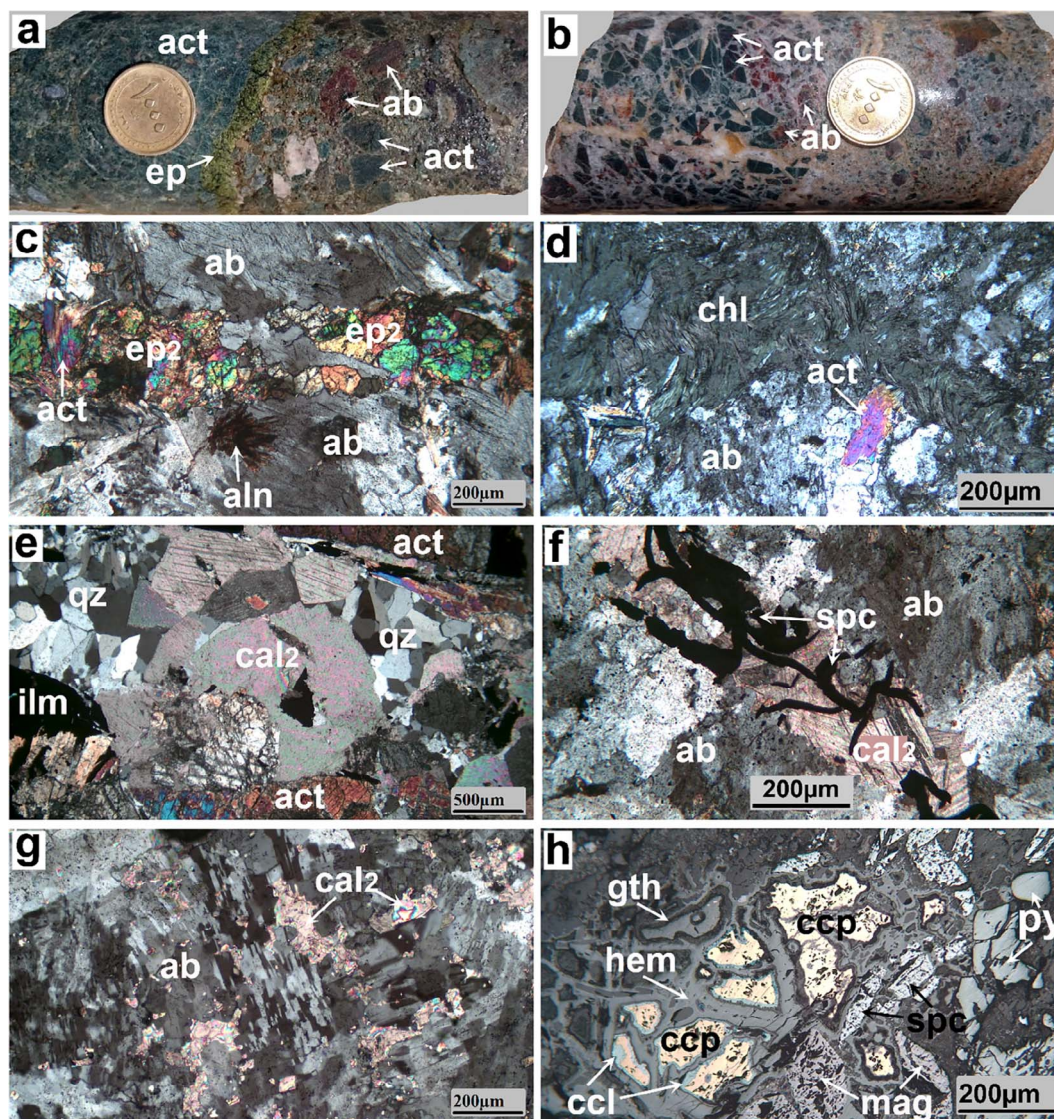


Fig. 20. (a, b) Photographs of the late stage breccias within the Saghand Anomaly 5: the breccias contain the fragments of actinolite and albite and various country rocks set into a matrix of epidote₂, hematite, clay minerals and carbonates; (c, d, e) the previous metasomatites intersected by the veins of epidote₂ (c), chlorite (d) and calcite₂ + quartz (e); (f) calcite₂ veinlet contains the specularite fibers that crosscuts the albite; (g) calcite₂ minerals are superimposed on albite during the late stage alteration and (h) deposition of specularite, replacement of magnetite₂ and pyrite by hematite and goethite, and replacement of chalcopryrite by hematite and chrysocolla in the late stage alteration. Abbreviations: ab = albite, ep = epidote, act = actinolite, aln = allanite, chl = chlorite, cal = calcite, qz = quartz, ilm = ilmenite, spc = specularite, gth = goethite, hem = hematite, ccp = chalcopryrite, py = pyrite, ccl = chrysocolla, mag = magnetite.

(Fig. 26). The minimum temperatures measured from stable isotopes vary from ~748° to ~766 °C for Na–Ca-metasomatism and 703° to 737 °C for potassic K–Ca-metasomatism (Table 4), and 350° to 400 °C for the veinlet-type late alteration stage (Fig. 27). These temperatures can be consistent with the temperature of a magmatic hydrothermal fluid. Based on the field and mineralogical observations, the Ca-alteration in the paragenesis sequence has occurred after the Na–Ca-metasomatism and K–Ca-metasomatism and prior to the late alteration stage, therefore, its formation temperature is placed between their formation temperatures. This data indicates the important role of magmatic waters in all stages of metasomatism and alterations in Saghand Anomaly 5. The lower temperature alteration assemblages (late alteration stage) are interpreted as shallow level alteration, in contrast to the Na–Ca-metasomatic, K–Ca-metasomatic and Ca-alteration assemblages, and might be related to hydrolytic alteration occurring due to the post-mineralization episode(s) of fluid circulation at shallower depths. As a result, it can be inferred that probably progressive evolution of a single source with magmatic source, contributed to the formation of Saghand Anomaly 5. Although several fluid

inclusion and isotopic studies that are carried out on several deposits in the Bafq district, such as Esfordi, Choghart, Se-Chahun, and Chardormalu (e.g., Daliran, 2002; Jami et al., 2007; Torab, 2008; Bonyadi et al., 2011; Taghipour et al., 2015; Heidarian et al., 2016) argue for an evolved magmatic hydrothermal origin (plutonic- and volcanic- related), modified by fluids involved in the later stages of mineralization, alteration, and recrystallization, but we have not found any evidence of non-magmatic fluids involved in the mineralization of Saghand Anomaly 5.

Most U ± REE deposits associated with alkali-metasomatism are hosted by granitic or felsic volcanic rocks with high U, Th and REE contents such as Lagoa Real in Brazil, Kurupung in Guyana, Timdrart in Morocco and Novoukrainska and Kirovograd in Ukraine (Cuney et al., 2012). At the Saghand Anomaly 5, the average values of U, Th and Th/U in granite are 3.66 ppm, 18.15 ppm and 5.87, respectively (Table 1). These values are approximately close to the average of these elements in crust granites (U = 4.8 ppm and Th = 17 ppm). In diorite-quartz diorite host rocks, the average values of U, Th and Th/U are 6.65 ppm, 8.39 ppm and 1.89, respectively. These values indicate that the leaching

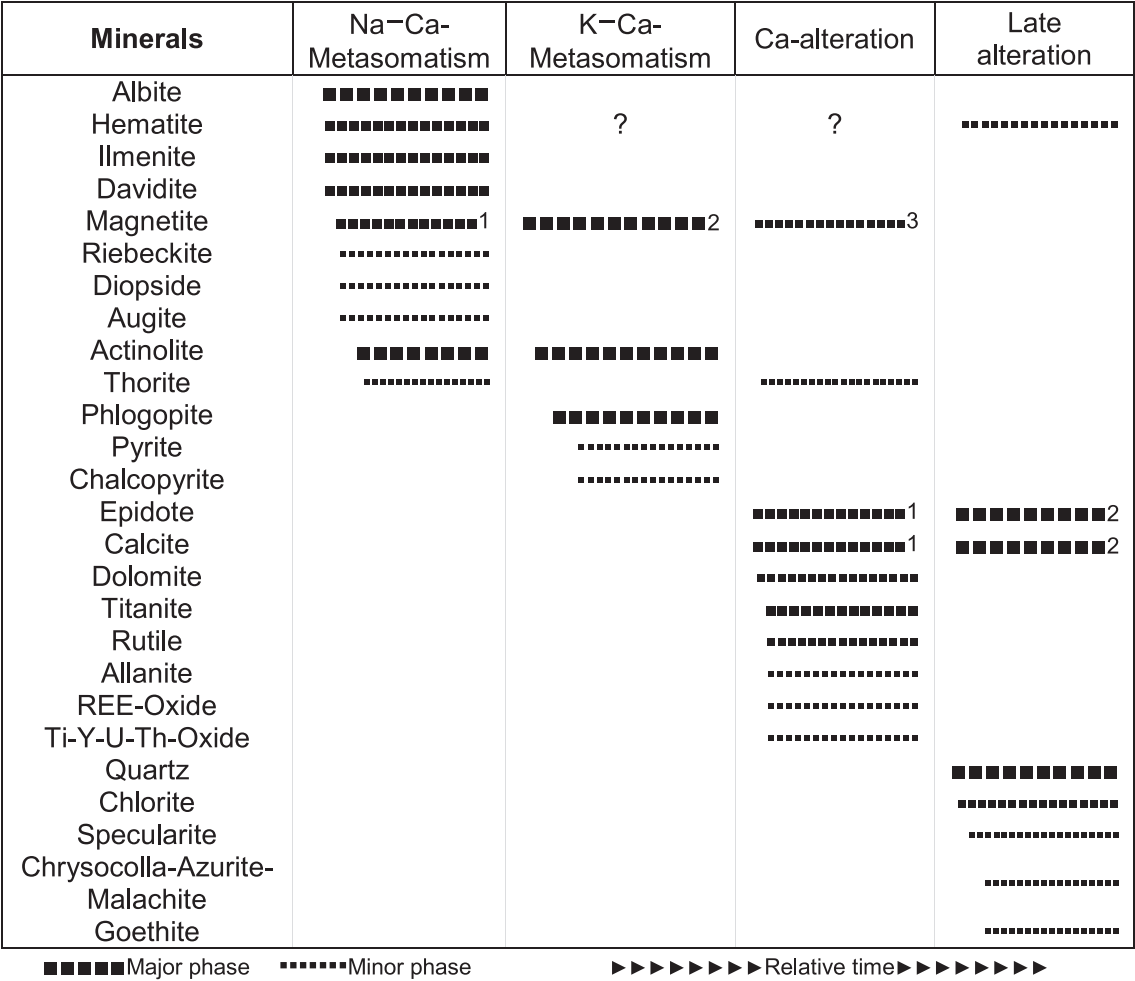


Fig. 21. Mineral paragenesis in metasomatites of the Saghand Anomaly 5. Numbers refer to mineral generation described in the text.

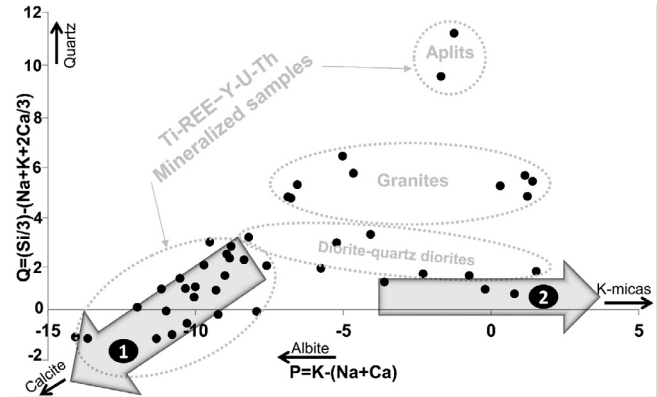


Fig. 22. Evolution of the quartz content (Q parameter) relative to K-feldspar and Na-Ca bearing mineral content (P parameter) in the Douzakh-Darreh Leucogranite, host diorite and quartz diorite and metasomatic zones associated with Ti-REE-Y-U-Th mineralization. Diagram is from Debon and LeFort (1988). Arrows show metasomatic stages: 1: Na-Ca-metasomatism; 2: K-Ca-metasomatism.

of elements from these rocks along the fluid migration pathways cannot be a very important source for mineralization. Nevertheless, the current relatively high Th/U ratios of granites compared to the bulk earth ratio ($Th/U = 3.9 \pm 0.1$) may indicate a U loss from these rocks. The mineralized metasomatites of the Saghand Anomaly 5 display relatively LREE-enriched patterns (Fig. 24a), which can reflect a felsic magmatic origin (cf. Marschik et al., 2000; Ray and Dick, 2002; Sillitoe, 2003;

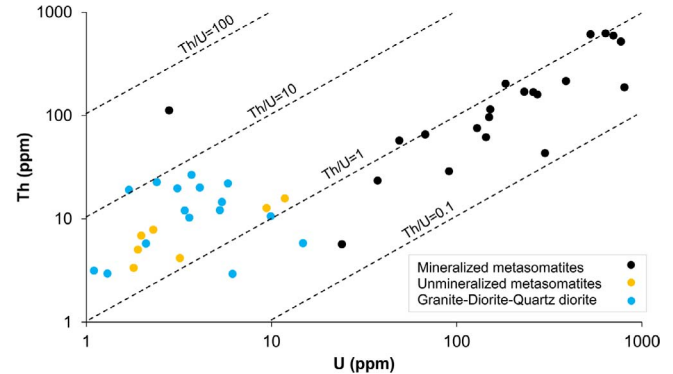


Fig. 23. Th versus U diagram for various rocks of the Saghand Anomaly 5.

Groves et al., 2010). The more detailed experimental data from Simon et al. (2004) demonstrated that significant amounts of Fe can be transported by a low density aqueous fluid in the magmatic hydrothermal environment. Saline, low-S fluids can carry many metals at high temperature and will precipitate Fe and other non-chalcophile elements on cooling, wall rock reaction, or mixing (Barton, 2014). The occurrence of REE deposits of indisputable hydrothermal origin, including the giant Bayan Obo REE deposit in China (Chao et al., 1992; Smith and Henderson, 2000), Gallinas Mountains deposit in New Mexico (Williams-Jones et al., 2000), and Karonge deposits in Burundi (Lehmann et al., 1994), leaves little doubt that, at least in some geological environments, the REEs are highly mobile. The Bayan Obo and

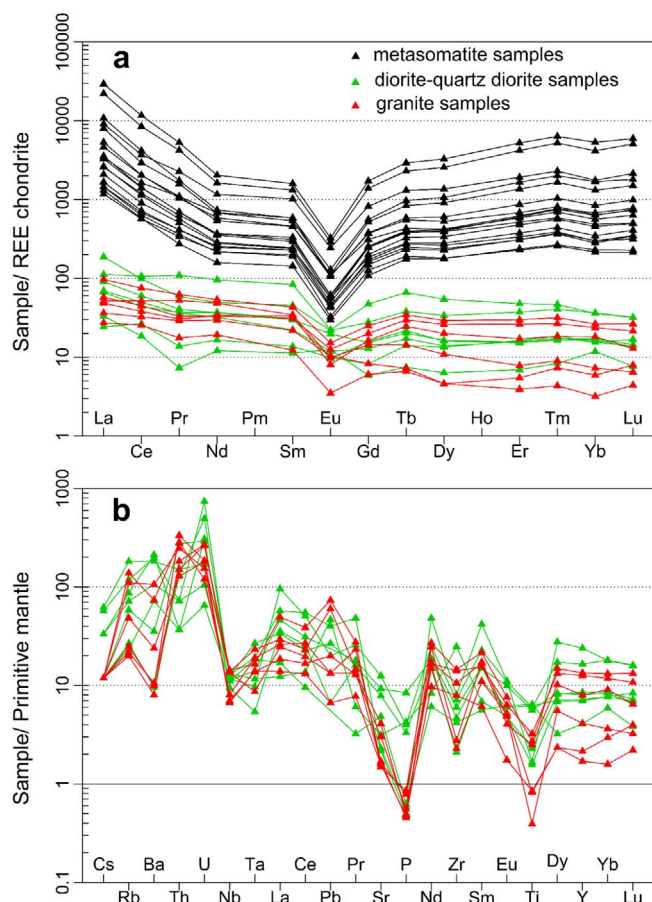


Fig. 24. Chondrite normalized (Nakamura, 1974) rare earth elements and incompatible elements distribution patterns (Spider diagrams) (McDonough and Sun, 1995) for the Saghand Anomaly 5 mineralized metasomatite, granite and diorite-quartz diorite host rocks.

Gallinas Mountains deposits are interpreted to have been formed between 300 and 400 °C and preliminary data for the Karonge deposits suggest that they were formed at > 420 °C (Williams-Jones et al., 2012). These hydrothermal REE deposits and other well-known hydrothermal deposits are all LREE-enriched and LREEs are more mobile hydrothermally than the HREEs (Chao et al., 1992; Williams-Jones et al., 2000). Based on the experimental investigations, the REEs are readily concentrated by hydrothermal processes, the REEs form stable complexes with fluoride and chloride ions at the high temperature (and likely sulfate, carbonate and phosphate), and the LREE complexes with these ligands are more stable than their HREE counterparts (Williams-Jones et al., 2012). The results of Mysen and Kushiro (1977) indicate that the rare earth elements (REE) would be partitioned strongly into fluid released by amphibole dehydration and that water could also readily leach REE from peridotite in the mantle wedge. In the Saghand Anomaly 5, therefore, the abundance of REEs in hydrothermal phases such as davidite, allanite-(La), Ti-Y-U-Th-Oxide and REE-Oxide (Table 2), REEs enrichment in metasomatized rocks relative to host rocks (Fig. 24a), the concentration of these elements in the development centers of the metasomatic zones (Fig. 3)- which are faults and brecciated zones- and the LREEs enrichment relative to HREEs in mineralized rocks (Fig. 24a) all indicate a significant mobilization and deposition of REEs, especially La, Ce and Y, by high temperature solutions.

Theoretical considerations suggest that trivalent REE and Y should exhibit strong, predominantly electrostatic complexing with “hard” ligands such as fluoride, sulfate, phosphate, carbonate and hydroxide (Williams-Jones et al., 2012). Complexing of these metals with chloride

and nitrate is moderately weak, and with ammonia and bisulfide is extremely weak to nonexistent (Wood, 1990b). In the Saghand Anomaly 5, the lack of S, F and P minerals suggests that sulfate, fluoride and phosphate complexes were insignificant. Meanwhile, concentrations of fluoride and phosphate are not expected to be high enough to increase uranium solubility significantly, considering the low concentrations of these ligands in arc magmas (Bailey and Ragnarsdottir, 1994). However, metasomatic fluids, generated by dehydration and decarbonation of amphibolite facies minerals in the subducting slab, are thought to be the most likely candidates for transporting and enriching the excess LILE (e.g., U and Th) and light rare earth elements (LREE) into the magmas (Bailey and Ragnarsdottir, 1994). Fluids generated by such dehydration reactions are believed to be of H₂O-CO₂ composition (Bailey and Ragnarsdottir, 1994). Evidence for such fluids are observed in Saghand Anomaly 5 as hydroxyl minerals (actinolite) in the Na-Ca-metasomatic stage and carbonate phases in the Ca-alteration stage. Considering the absence of carbonate phases paragenesis with davidite in Na-Ca-metasomatic stage and that with increasing pressure and temperature the hydroxide species become more soluble than the carbonate species (Bailey and Ragnarsdottir, 1994), therefore, hydroxide complexation appears to provide a mechanism for REE, U and Th enrichment under conditions of high oxygen fugacity in the Na-Ca-metasomatic stage. Th(OH)₂²⁺ and (UO)₂(OH)₂²⁺ can be some of the transporting species (Bailey and Ragnarsdottir, 1994). Considering that carbonate phases (dolomite and calcite) are paragenesis with allanite-(La), Ti-Y-U-Th-Oxide and REE-Oxide at the Ca-alteration stage at the Saghand Anomaly 5, it seems most likely that U, Th and REE were transported together in a hydrothermal fluid as soluble carbonate complexes during this stage. Typical forms would be (UO₂(CO₃)₃)⁴⁻, Th(CO₃)_n⁽²ⁿ⁻⁴⁾⁻ and (REE(CO₃)₃)³⁻ (McLennan and Taylor, 1979; Titayeva, 1994). REE, U and Th carbonate complexes are best formed under alkaline (ph ≥ 7.5) and oxidizing (Eh ≥ 0.2 V) conditions with high temperatures (McLennan and Taylor, 1979). If soluble carbonate complexes are controlling REE distribution in mineralized metasomatites of Saghand Anomaly 5, the Eu²⁺ would be expected to behave very differently than REE³⁺. This should result in anomalous Eu behavior in the REE patterns; this clearly occurs at the Saghand Anomaly 5 (Fig. 24a).

According to low background values of the Ti, REE, U and Th in the host rocks (Table 1), the lack evidence of the unusual magmas such as carbonatite or alkaline magmas at the area, and the mobility potential of REEs, U and Th in the magmatic hydrothermal solutions, especially LREEs, we propose that the most likely source of supply and transportation of these elements in the Saghand metasomatic system is the early Cambrian calc-alkaline granitic magma. In addition, the calc-alkaline plutonic rocks only represent a significant U, Th and REEs source when the structure of U-Th-REEs-bearing accessory minerals (such as the zircon, apatite, monazite, titanite, xenotime, thorite, uranophane and allanite) is destroyed during later fluid circulation events as a result of alpha-recoil during U/Th decay (metamictization) (Cuney, 2014). Significant metamictization of the U-Th-bearing silicates typically requires a time laps of 200 Ma. Therefore, due to the simultaneity of mineralization with calc-alkaline magmatism in the Saghand area, there was not enough time for metamictization and leaching of these elements from the Douzakh-Darreh granite. Accordingly, the fluids differentiated from Douzakh-Darreh granite have been entered into fault and crushed zones in a tectonically active regime of marginal continental arc. Due to reaction of the high temperature fluid with the host rocks, the ratios of Na⁺/K⁺ and Na⁺/H⁺ or alkalinity of fluid in equilibrium with feldspars of host rock are elevated (Cuney et al., 2012). Increased fluid alkalinity causes albite precipitation and quartz dissolution. This basically alkaline and high temperature hydrothermal fluid is a suitable environment for the solubility and transfer of U, Th and REE as carbonate complex (McLennan and Taylor, 1979). The paragenesis of alteration minerals such as micro-hematite in micro fractures of albite (Fig. 8 h) and crystallization of the Fe³⁺-bearing silicates (i.e.

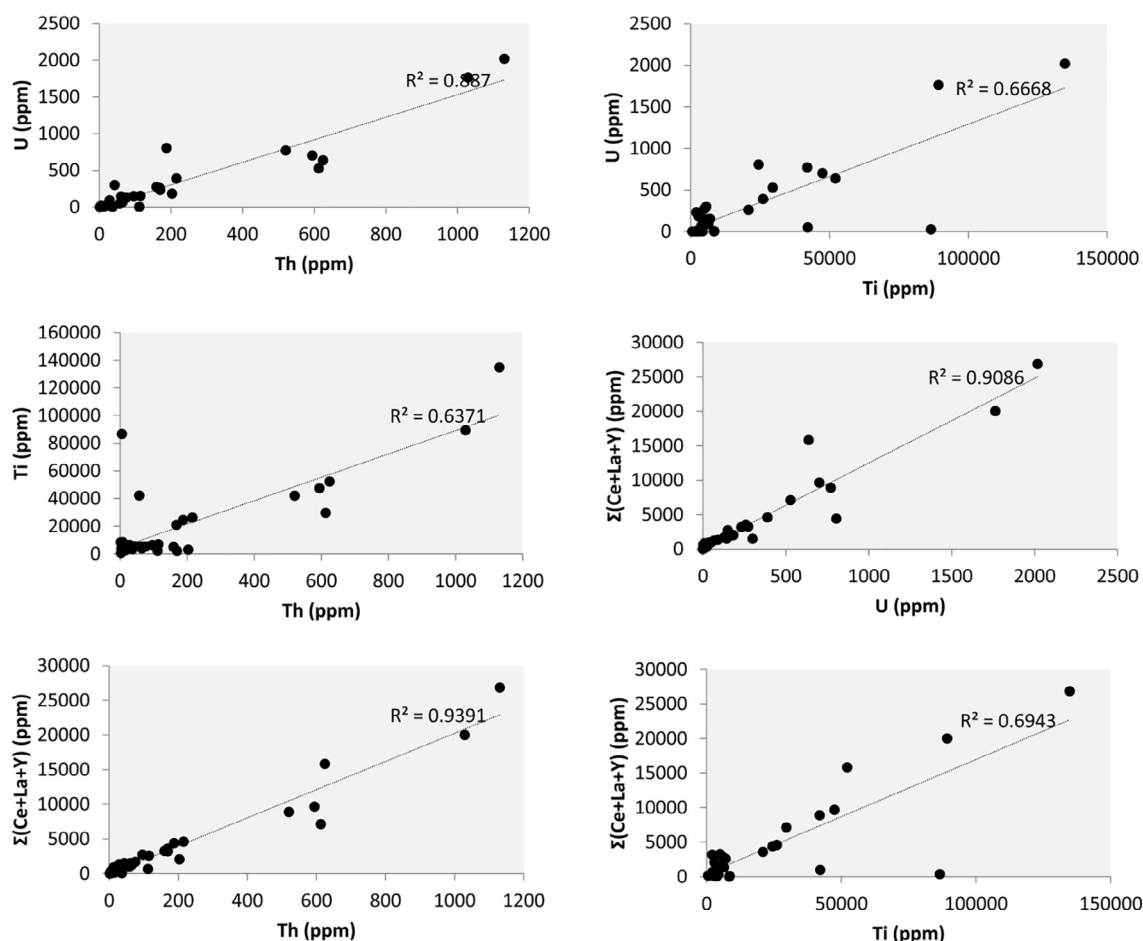


Fig. 25. Correlation between Ti, REE, Y, U and Th elements in mineralized metasomatic rocks of the Saghand Anomaly 5.

Table 3

Hydrogen, oxygen and carbon isotopic values and water concentrations in various phase and stages within Saghand Anomaly 5.

Sample	Phase	Stage	δD ‰ SMOW	H ₂ O wt%	$\delta^{18}O$ ‰ SMOW	$\delta^{13}C$ ‰ PDB
PS-1	Phlogopite	K–Ca-	–59.3	3.99	6.25	
PS-2	Phlogopite	metasomatism	–66.5	4.01	6.37	
PS-3	Phlogopite		–62.1	4.09	6.72	
AS-1	Actinolite	Na–Ca-	–64.8	1.92	7	
AS-2	Actinolite	metasomatism	–70.4	2.15	7.39	
AS-3	Actinolite		–68.5	2.23	6.36	
AS-4	Actinolite		–64.4	1.36	6.45	
AS-5	Actinolite		–66.5	0.37	6.01	
MS-1	Magnetite ₂	K–Ca-			2.39	
MS-2	Magnetite ₁	Na–Ca-			3.45	
MS-3	Magnetite ₁	Na–Ca-			3.34	
MS-4	Magnetite ₁	Na–Ca-			3.14	
MS-5	Magnetite ₂	K–Ca-			2.33	
CS-1	Calcite ₂	Late alteration			10.64	–5.53
CS-2	Calcite ₂	Late alteration			10.54	–5.44

riebeckite), the negative Eu anomaly relative to trivalent REEs in mineralized zone, and transform of U as U⁶⁺ all indicate that the mineralizing solutions during the Na–Ca-metasomatic stage would have been oversaturated in sodium and were oxidized. During deposition of the

albite and hematite, the alkalinity and oxidation degree of the fluid are gradually dropped. Alternatively, the activity of CO₂ effectively decreased by the lowering of total pressure on entrance of the ore-bearing solutions into fault and shear zones. Therefore, lowering of pH, Eh and carbonate activity are all efficient mechanisms for the instability of the complexes. Ti in the hydrothermal fluid forms the titanium hydroxide, which is a strong adsorbent surface for a large amount of U, Th and REEs from the fluid (Zhao, 2005). After the stopping hydrothermal activity, the sub-stable titanium hydroxide changes into the stable Ti-minerals (davite and ilmenite). When sodium is gradually removed as albite from the hydrothermal fluid(s), calcium activity in the fluid increases and the pH decreases, and due to the deposition of hematite and Fe³⁺-bearing silicates, the Eh decreases. Therefore, the conditions for actinolite deposition simultaneously or after the albite precipitation are provided. Due to the exit of sodium ions from fluids and increased potassium activity, K–Ca-metasomatism and magnetite₂ mineralization have occurred. In the following stage of hydrothermal fluid evolution, the infiltration of Ca-rich fluids cause the formation of Ca and Ca-bearing minerals such as calcite₁, dolomite, titanite, epidote₁ and allanite-(La) as well as Ti–Y–U–Th-Oxide and REE-Oxide at the Ca-alteration (Figs. 14, 15 and 17). These fluids also altered the ilmenite and davite minerals and formed the magnetite₃ and titanite at this alteration stage (Figs. 12 and 13). Finally, as a result of decreasing of hydrothermal fluid temperature and changing in the chemical composition of the fluid, the late alteration stage occurred and veinlets of epidote₂ ± chlorite ± calcite₂ ± quartz were contemporaneously formed.

Table 4
Oxygen geothermometry, Saghand Anomaly 5.

Stage	Sample No.	$\delta^{18}\text{O}$ mineral pair	A factor	Temperature ($^{\circ}\text{C}$)
Na–Ca-metasomatism	AS-4 MS-3	act (6.45‰)- mag (3.34‰)	3.36 (Bilenker et al., 2016)	~766
Na–Ca-metasomatism	AS-3 MS-4	act (6.36‰)- mag (3.14‰)	3.36 (Bilenker et al., 2016)	~748
K–Ca-metasomatism	PS-2 MS-5	phl (6.37‰)-mag (2.33‰)	4.13 (Chen et al., 2010)	~737
K–Ca-metasomatism	PS-3 MS-1	phl (6.72‰)- mag (2.39‰)	4.13 (Chen et al., 2010)	~703

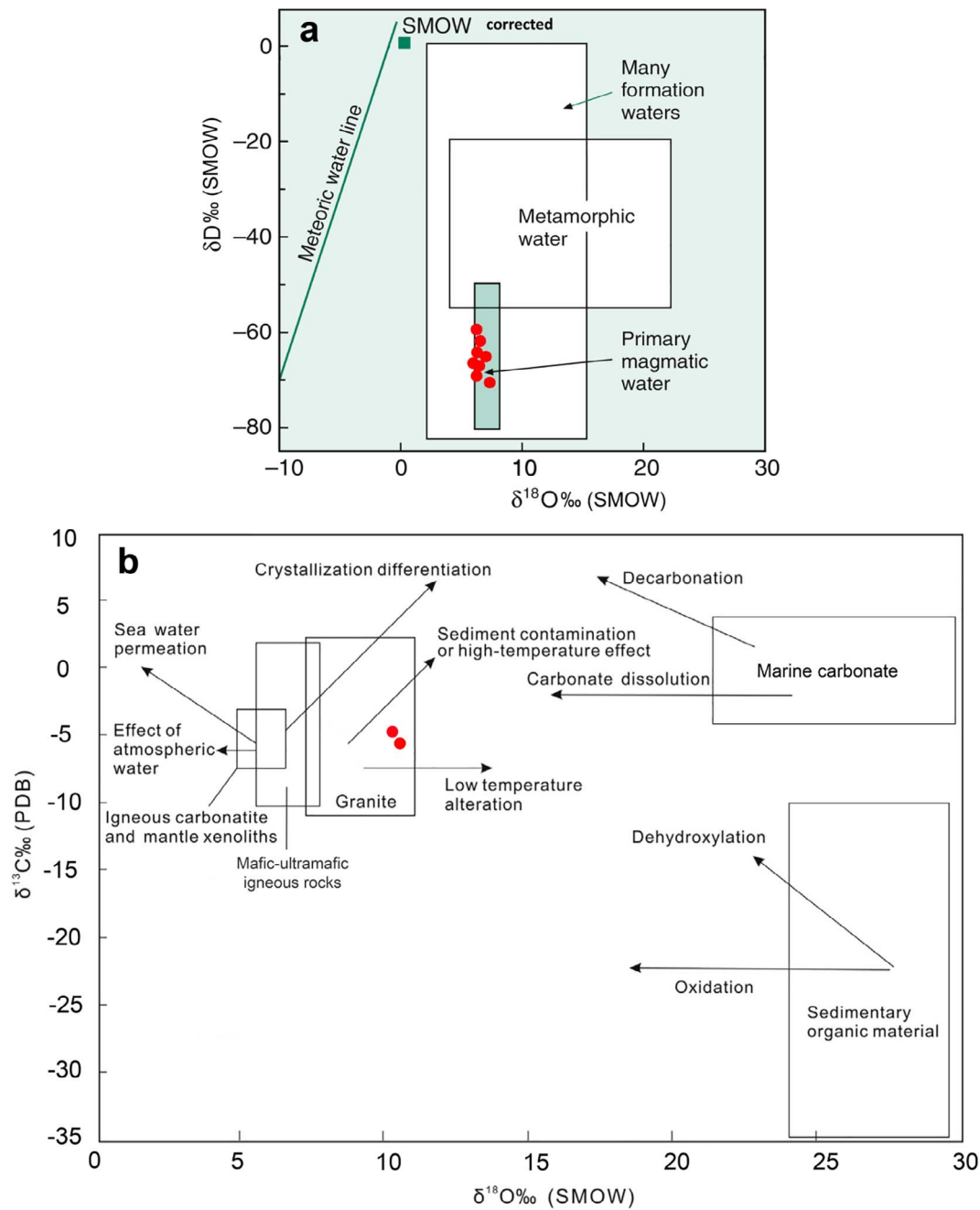


Fig. 26. (a) O–D isotopic data for the actinolite and phlogopite within the Saghand Anomaly 5 with respect to the meteoric water line and the isotopic composition of metamorphic and magmatic fluids (fields from Hoefs, 2009); (b) The $\delta^{13}\text{C}_{\text{PDB}}$ and $\delta^{18}\text{O}_{\text{SMOW}}$ values of calcite₂ from the Saghand Anomaly 5. References for the fields of major carbon reservoirs and other deposits are from Liu et al. (1998).

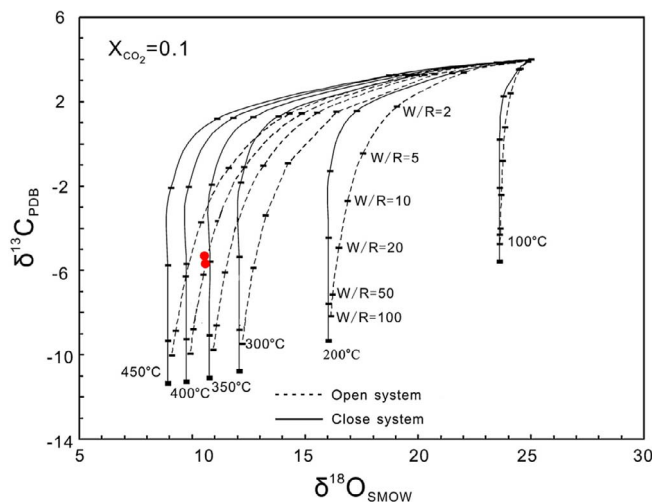


Fig. 27. The location of calcite₂ of the Saghand Anomaly 5 in the temperature range of 350 to 400 °C on the $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ - XCO_2 diagram of Zheng and Hoefs (1993).

7.2. Metasomatic evolution

Based on field evidence and microscopy studies, four main stages of metasomatism and alteration with continuous evolution have been distinguished at the Saghand Anomaly 5, including: 1) Na–Ca-metasomatism, 2) K–Ca-metasomatism, 3) Ca-alteration, and 4) epidotization $\pm$ chloritization $\pm$ silicification $\pm$ carbonatization late alteration that is associated with brittle deformation and superimposed on the earlier metasomatic zones. Contrasting mineral parageneses developed during these stages in response to decreasing P–T conditions and possibly variations in the chemistry characteristics of the solutions. However, the variations in the mineral parageneses can also reflect the nature of the host rocks subjected to metasomatism. The Na–Ca-

metasomatism and Ca-alterations were pervasive, due to infiltration of hydrothermal fluids along fractures and shear zones (Fig. 3), but meanwhile K–Ca-metasomatism is limited to the margin of the iron orebodies and tends to occur at depth (Fig. 3). Na–Ca-metasomatic zones are generally enriched in REE, Y, U and Th as compared to the host rocks (Fig. 11a, b), but K–Ca-metasomatic zones are usually associated with magnetite₂ mineralization (Fig. 6b). Similar metasomatic events have been already described in several other occurrences of iron deposits in the Bafq-Saghand area with an initial stage of Na–Ca-metasomatism followed by a K–Ca-metasomatic stage (Torab and Lehman, 2007; Heidarian et al., 2017).

Increasing volatile content during crystallization of magma enriches the melt in the albite content resulting in transfer of K from fluid to the wall rocks (potassic alteration), conversely when volatiles are lost from the system, the residual melt becomes enriched in orthoclase, and Na in the fluid is transferred to the wall rocks (sodic alteration) (Pirajno, 2009). Therefore, it is impossible for potassium and sodium in the nature to occur simultaneously in the hydrothermal fluids. In other words, only one of these alterations occurs at any time and never occurs simultaneously.

At the Saghand Anomaly 5, albitization occurred more under ductile conditions, as evidenced by the granoblastic recrystallization of albite and bending of albite twins (Fig. 8e) in response to ductile deformation (Tullis and Yund, 1985) at high temperatures ($T > \sim 748^\circ\text{C}$). Continuing the Na–Ca-metasomatic stage, temperature decreased and deformation became brittle with development of fracturing and brecciation of albite and formation of actinolite as fracture filling and vein/veinlet (Fig. 8b and c.).

During the Na–Ca-metasomatic stage, ilmenite and davidite ore minerals together with albite, riebeckite, diopside, augite and actinolite were deposited (Fig. 11a, b). At the Ca-alteration stage, as a result of infiltration of carbonate and Ca-rich solutions, ilmenite and davidite were altered to magnetite₃ and titanite (Figs. 12 and 13), as well as rutile, allanite-(La), Ti–Y–U–Th-Oxide and REE-oxide were formed

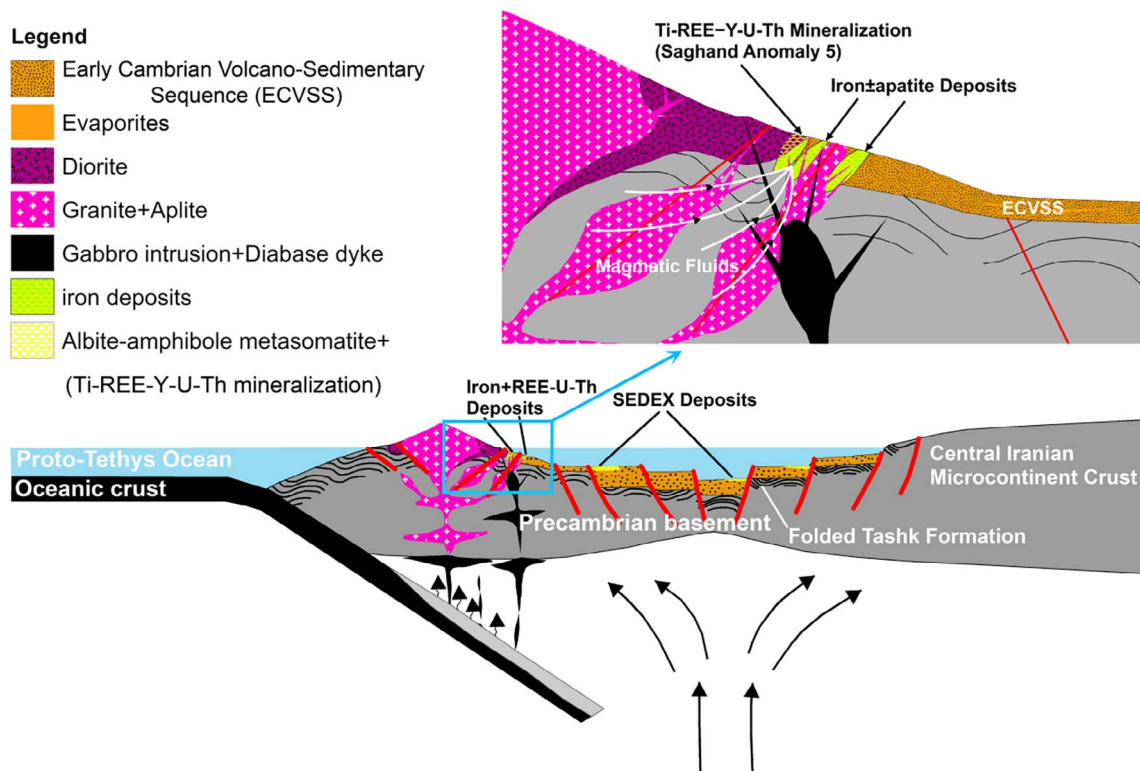


Fig. 28. A possible schematic reconstruction model illustrating the relationships between magmatism and iron-(Ti-REE-Y-U-Th) mineralization at the Saghand Anomaly 5 (modified from Rajabi et al., 2015).

along with epidote₁, calcite₁ and dolomite (Figs. 14, 15 and 17). The mineralizing solutions during this stage would have been oversaturated in Ca and were still oxidizing as evidenced by crystallization of the Fe³⁺-bearing minerals such as magnetite₁, allanite-(La), epidote₁ and trivalent rare earth elements oxide.

7.3. Metallogenic model

The tectono-magmatic evolution of Central Iran has been debated for a long time, especially the Bafq-Saghand district and associated iron deposits. A common structural model proposed for the Early Cambrian tectonics and magmatism in the Bafq-Saghand district is crustal extension associated with intra-continental rifting (Berberian and King, 1981; Talbot and Alavi, 1996; Samani, 1998). It has been proposed that this rifting event formed an aborted rift associated with alkaline magmatism, including alkaline volcanic rocks and alkaline granites (e.g., Samani, 1988, 1998; Daliran, 1990, 1999, 2002; Förster and Jafarzadeh, 1994). However, based on trace element characteristics of granitic intrusions and felsic volcanic rocks, Ramezani and Tucker (2003) suggested that the Kashmar-Kerman volcano-plutonic arc (including Saghand district) represents an active continental-margin environment in relation to subduction of Proto-Tethyan Ocean beneath the Central Iranian microplate in the Early Cambrian.

According to Samani (1998), tectono-magmatic activity and mineralization in Saghand district is rift-related and rifting was associated with carbonatic metasomatism and by intrusion of granitic masses of metasomatic origin that were accompanied by a regional Na-metasomatism. But a synthesis of key geological and geochemical evidence in the Saghand Anomaly 5, such as extensive felsic volcanic associations, diorite and quartz diorite subvolcanic rocks, and leucocratic granitic rocks (Fig. 2), calc-alkaline affinity of granite and host rocks (Fig. 4b), enrichment of U, Th and LREE compared to Nb, Ta, P, Ti and HREE in granite and host rocks (Fig. 24) and other geochemical data all indicate that tectono-magmatic setting and its associated Fe and Ti-R-EE-Y-U-Th mineralization could be related to a subduction zone and continental-margin arc setting magmatism (Fig. 28).

Many REE-U-rich provinces are related to evolved felsic igneous rocks intruded into the crust, not only anorogenically, but also during the final stages of orogenesis (Plant et al., 1999). During the late orogenic stages, ductile shear fault zones controlled the site of emplacement of felsic magmatic provinces (Bonin, 1987). Pressure release caused by the faulting can produce a partial melting in the deep zones of the thickened orogenic lithosphere. Furthermore, the partial melting of the lithospheric mantle above the subducted slab is also supported by the dehydration of the latter. The interaction between fluids generated during this dehydration and overlaying mantle material would be responsible for the Th, U, and REE enrichment in magmas (Keppler, 1996). When submitted to fractional crystallization processes, such magmas eventually evolve into U-Th rich felsic lithotypes and trace and rare earth elements-rich fluids are released (Chaves, 2013).

Given the results mentioned above, we suggest that the mineralogical and geochemical data are best interpreted in terms of a magmatic origin for the mineralization in the Saghand Anomaly 5. Furthermore, stable isotopic data are consistent with a magmatic source for ore-forming fluids. Therefore, the probability that the Saghand Anomaly 5 were related to basinal brines from an evaporitic source is extremely small. However, it can be inferred that high temperature fluids differentiated from calc-alkaline granitoid magma would have ascended along faults and fractures, and after the proximity to the surface and the fluid/rock reaction, caused the metasomatism, hydrothermal alteration and mineralization (Fig. 28). In this model, granitoid magma and diorite-quartz diorite rocks serve as the source and hosts for mineralization, respectively (Fig. 28).

Acknowledgements

This study is part of a PhD thesis by S. Deymar at the Shahid Beheshti University, Iran. This research was also supported by Atomic Energy Organization of Iran (AEOI) and Iranian Mines and Mining Industries Development & Renovation (IMIDRO).

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at <http://dx.doi.org/10.1016/j.oregeorev.2018.01.008>.

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