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GEOCHEMISTRY OF MID-UPPER EOCENE ALKALINE VOLCANIC ROCKS, ZIARAN VOLCANIC BELT, SOUTH OF CENTRAL ALBORZ MOUNTAINS, NORTH OF IRAN

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ABSTRACT

The Ziaran volcanic Belt (ZVB), North of Iran contains a number of intra-continental alkaline volcanic range situated on South part of central Alborz Mountains, formed along the localized extensional basins developed in relation with the compressional regime of Eocene. The mid-upper Eocene volcanic suite comprises the extracted melt products of adiabatic decompression melting of the mantle that are represented by small volume intra-continental plate volcanic rocks of alkaline volcanism and their evaluated Rocks with compositions representative of mantle-derived, primary (or near-primary) melts. Trace element patterns with significant enrichment in LILE, HFSE and REEs, relative to Primitive Mantle. Chondrite-normalized of rare earth elements and enrichment in incompatible elements and their element ratios (e.g. LREE/HREE, MREE/HREE, LREE/MREE) shown these element modelling indicates that the magmas were generated by comparably variable degrees of partial melting of garnet lherzolite and a heterogeneous asthenospheric, OIB mantle sources.

***Keywords*:** Ziaran area; Iran; garnet lherzolite; partial melting.

Introduction

There are outcrops of alkaline volcanic rocks of the upper middle Eocene age in the southern part of central Alborz in Ziaran region. The prominent feature of the rocks is the presence of mafic minerals and the layered structure of the sequences of volcanic rocks which are outcropped unconformably on the middle Eocene pyroclastic rocks of Karaj formation, sometimes with a fault boundary.

Alborz Mountains in northern Iran and southern Caspian Sea with an eastern-western direction form a relatively zig-zag path. Alborz Mountains are part of the Northern Alpine-Himalayan orogeny belt in Western Asia which is limited to the Caspian block from the north and the central Iranian plateau from the south. In terms of orography, the west border of Alborz extends to Lesser Caucasus and its external border extends to *Paropamisus* Mountains in Afghanistan.

Paleogene magmatism has been widely extended in Iran with outcrops in different parts of Iran, including Central Iran (Urmia-Dokhtar magmatic strip, southern margin of the Alborz orogeny strip, the Western Alborz-Azerbaijan, the Lut block and the north of Lut .This magmatism extends beyond Iranian land and is also found in neighboring regions such as Afghanistan and Turkey, and even in the southern margin of Eurasia before paleogenesis. Magmatism has been carried out over a broad period of time from Triassic to Quaternary.

Due to abundance of tertiary volcanic and pyroclastic rocks in the southern slopes of Alborz, Alborz was considered as part of the large Caucasus-Turkey Stalagmite in the first European-made world map. However, the presence of the same magmatic rocks in other parts of Iran, especially further findings from the geology of Iran confirmed that most lithostratigraphic units in Alborz and Central Iran are similar in terms of facies and formation conditions. Accordingly, Alborz can be considered as the marginal folds of Central Iran that the collision of Iran and Turan plates and its consequences have played a key role in its formation. Alborz is similar to Central Iran, especially on the southern slope, with some differences on the northern slope (Stocklin, 1968).

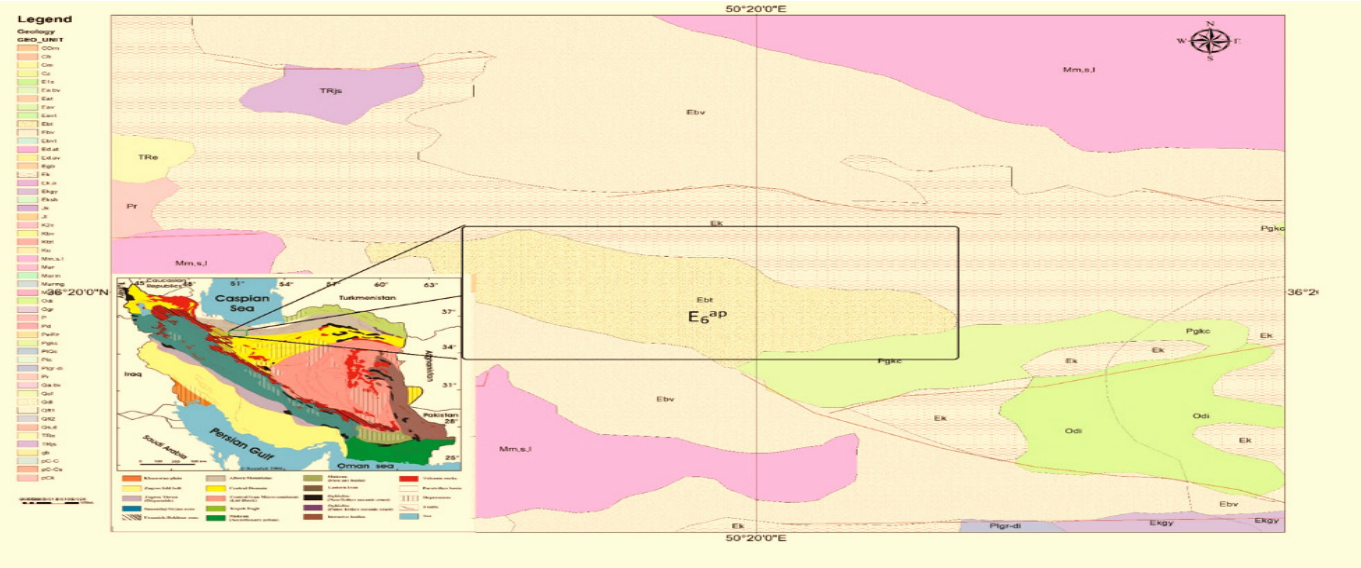


Fig. 1: The location of andesite-basaltic rocks of middle Eocene age in the Central Alborz and E6ap Alamut area in the major structural-sedimentary zones of Iran (Geological Survey of Iran).

In most areas of Alborz, Paleozoic-Middle Triassic sediments are conformable despite the lack of many hiatus indicating land-based tectonic movements. Although the events caused by the dynamic and active continental marginal collision of Turan with the static Alborz continental margin caused formation of thrust faults and upduction of the ancient Tethys oceanic complexes on the northern edge of Alborz in the Middle Triassic simultaneously with the former Cimmerian orogeny event, the first Alpine orogeny occurred in the Paleocene simultaneously with the Laramide orogeny event. This was associated with a thrust faulting, folding, and the rise of intermontane sedimentary basins, accumulation of clastic rocks simultaneously with orogeny event and pre-drought migration toward the south. The next orogeny occurred at the beginning of the Oligocene with consequences such as magmatism, rise of the earth's surface of water and the expansion of intermontane areas. The last phase of Alpine orogeny has taken place in the late Pliocene or early Pleistocene resulting in faulting, thrusting, elevation, and the current feature of Alborz.

The old formations of Alborz are located conformably and Cretaceous carbonates and older volcanic rocks in the southern-central Alborz have been folded. However, Paleocene-Eocene conglomerate (Fajan Formation) and Ziarat Formation (Eocene nomolithic limestone) are located uncorformably and form Karaj (Eocene) formation throughout northern Iran (Stocklin and Sotoudehnia, 1977). The Cretaceous sequence to Eocene implies the transition from a static tectonic period during Jurassic and Cretaceous to a regional thrust in the late Cretaceous to the Paleocene (Stocklin, 1974; Berberian and King, 1981). Karaj formation includes calc-alkaline volcanic units, volcanoclastic units and shale (Barberian and Barberian, 1981; Allen *et al.*, 2003). Karaj formation is uncorformably covered by shallow marine rocks and basalt conglomerate of Middle Eocene - Former Oligocene age. This sequence is developed with Oligocene and Miocene rocks in large parts of central-southern Alborz and central Iran (Guest, 2004).

Alborz Mountains have been formed by Alpine orogeny movements during the Tertiary period with a west-east direction and separate the coastal and mountainous plains of Alborz from the internal parts of Iran. Faults have played a significant role in various tectonic evolutions of Albroz (deformation, folding, magmatism, etc.). Most fissure volcanic eruptions in the central Alborz reached the earth surface through faults and openings and the presence of various reverse faults and thrusts has played a significant role in the central Alborz deformation. In other words, the current deformation of the central Alborz is more related to the reverse faults and thrusts and less related to strike-slip faults so that part of a fault acted in thrust mode while the other part in the extensional mode. This also affects the central Alborz magmatism. In general, Laramide orogeny event has both thrust and extensional properties. In the thrust phase, meso-Tethys closure has begun resulting in the formation of Iranian colorful mixes and thrusting them on the edges of continents and the placement of intrusive bodies, metamorphism and folding. The extensional phase of the Laramide event is a kind of emancipation after compression which occurred in the Paleocene-Eocene and its peak in the middle Eocene caused the formation of basic to intermediate alkaline volcanic rocks of the Eocene-Quaternary age.

The Upper middle Eocene volcanic strip of Ziraran area is located in the southern part of the 1/100000 Lankaran sheet with a composition of basic to intermediate volcanic units. It is extended on the volcanoclastic units of Karaj formation and exhibits most outcrops in the middle-southern parts of this sheet. There is a close connection between the placement of this volcanic unit and thrust faults (Fig. 1).

2. Petrography

In petrographic studies of volcanic rocks in the study area, the rocks are divided into olivine basalt, basalt, and basaltic andesitic, andesitic basaltic and andesitic rock ranges. The basaltic rocks are mainly composed of clinopyroxene, plagioclase, olivine and dark minerals as the minor minerals. The texture of the rocks is mainly porphyritic with a glassy-microgranular matrix (Figs. a, b, c). The andesitic rocks is mainly composed of a porphyritic texture with a microlithic matrix and its main mineral consists mainly of andesine, pyroxene and olivine plagioclase as the minor minerals (Fig. d, e, f).

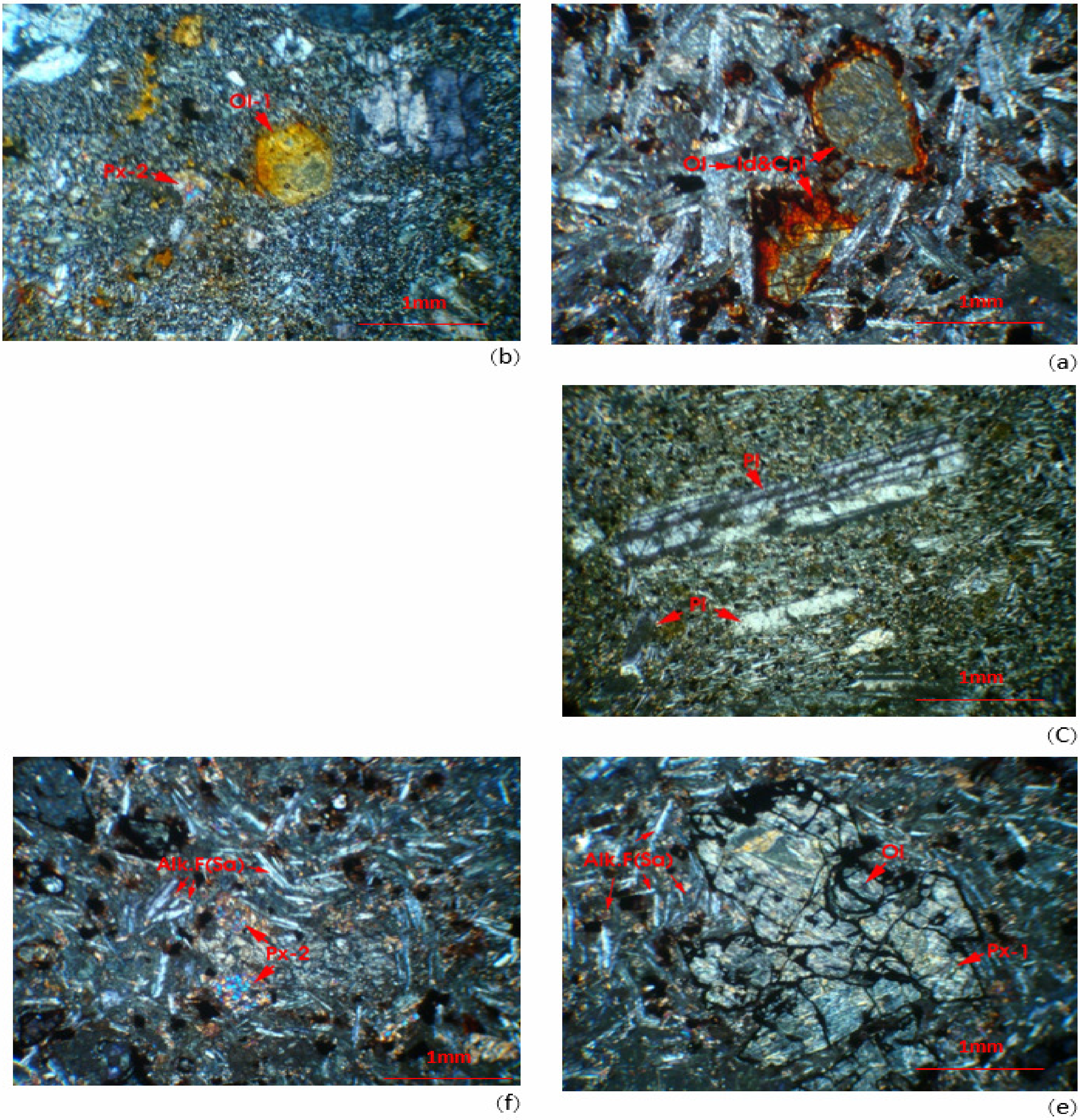


Fig. 2: Microscopic images of volcanic rocks of the middle upper Eocene age in Ziaran: (a, b, c) olivine basalt, olivine (with Iddengsite edges), clinopyroxene (Cpx), titaniaaugite (Ti, Aug), plagioclase (Pl) (andesine), dark minerals (plane-polarized light, opq), (c, e, f)) andesite, plagioclase (Pl) (andesine), olivine (olv), pyroxene (Px) (plane-polarized light)

3. Geochemistry of major and minor elements

Given the outcrops and extension of the volcanic sequence in the Ziaran area and the 1/100000 map of Shokran (Eb6 volcanic unit) of the upper Middle Eocene age on the Karaj Formation, 15 samples of rock units were collected from the volcanic sequence sections of the ZVB area (Table 1). Since some samples contain significant amounts of LOI, the original oxides were calculated on the Volatile Free basis by eliminating this parameter from the analysis list.

Figures 4 and 5 show the naming of ZVB rocks using the TAS diagram and changes in incompatible elements.

The total alkaline versus silica (Lebas *et al.* 1986) shows that most samples range from basalt to trachybasalt and all samples are located in the alkali basalt range (Fig. 2a). Some samples are considered more alkaline and located in the tephrite-basanite range. The total alkaline versus silica (Cox *et al.*, 1979) shows that most samples are located in the range of alkaline and basalt rocks and some evolution rocks are located in the Hawaiite range. In this diagram, the boundary between the alkaline and subalkaline ranges is adopted from Macdinald & katasura (1964) (Fig. 2b).

The De la Roche *et al.*, 1988 (Fig. 2c) graphs are also used in naming ZVB rocks. According to this diagram, the rock samples are located in the alkali basalt, tephrite, Hawaiite and lati-basalt range.

Pearce (1996) made some changes in the Zr / TiO_2 * 0.0001 versus Nb/Y diagram (Winchester & Floyd 1977). In this diagram, samples are located in the subalkaline basalt to andesite range (Fig. 2d). In the sub-alkaline region, Nb/Y equals 0.6 (Winchester & Floyd 1977, Pearce 1996). This can be attributed to melting of a certain mineral of a heterogeneous mantle with a high Y.

As can be seen, in most cases, the geochemical nomenclature of total alkaline versus silica and other diagrams for volcanic rocks are clearly consistent with nomenclatures based on petrographic studies. Comparing TAS diagrams in Fig. 2, it is obvious that alteration and metasomatism did not cause a severe alkaline displacement in the samples. In the microscopic sections, the dominant phase of alteration includes chlorination and serpentinization and argillization is less observed. The recent alteration was associated with significant displacement of alkali elements. Due to the chemical composition of chlorite, which does not contain Na and K, one can trust the nomenclature in the TAS diagram.

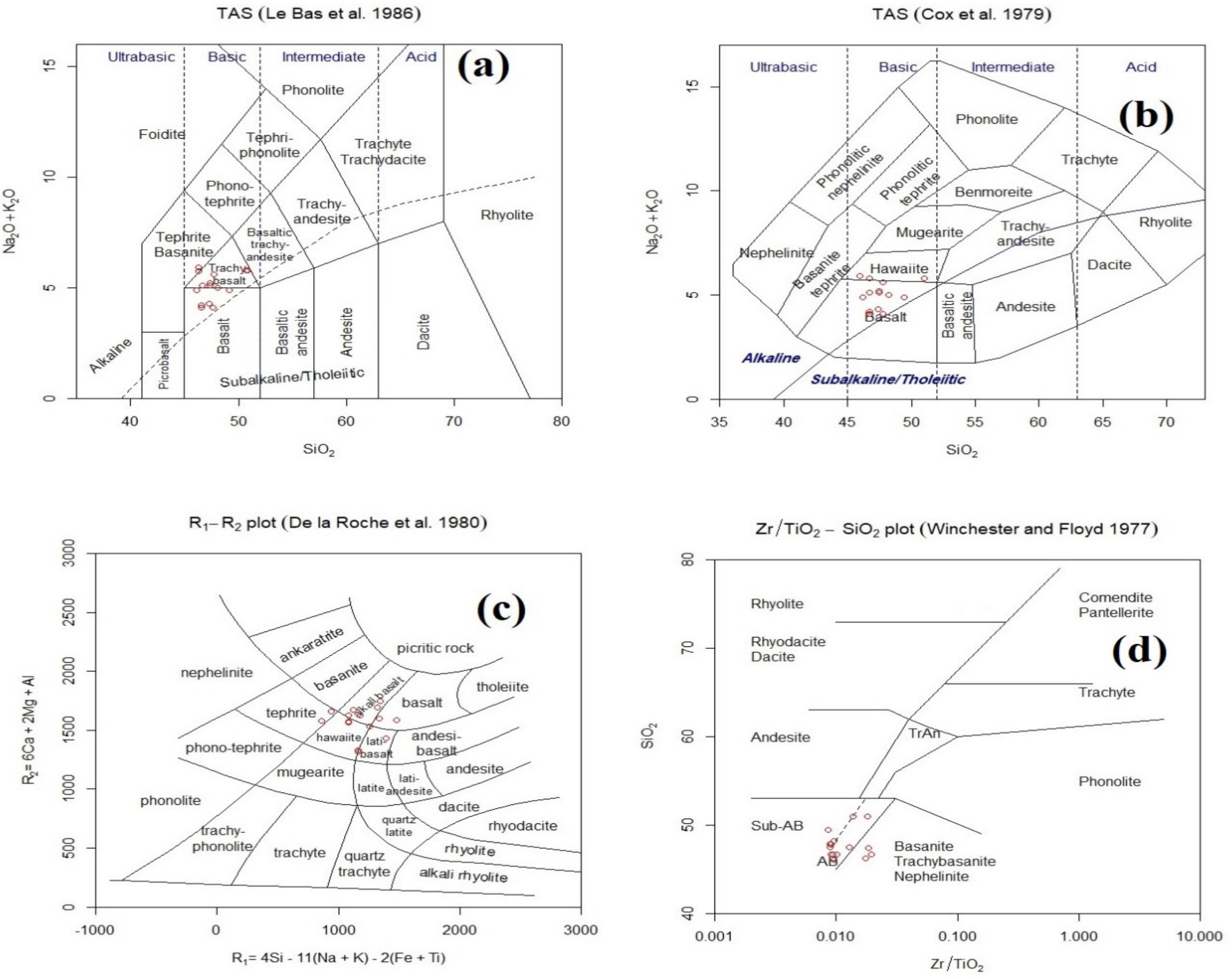


Fig. 3: Chemical classification of volcanic rocks in Ziaran based on Le Bas *et al.* (1992); Cox (1979); Winchester & Floyd (1977), Pearce (1996).

Multi-elemental graphs of incompatible elements, known as spider diagrams, are used to understand the frequency of rare elements in basalts relative to normalized references. These graphs are also widely used for understanding tectonic regimes and their magma evolutions (Wood *et al.*, 1979; Thomson 1982; Wilson 1989; Pearce, 1983).

The normalized pattern of the frequency of incompatible rare elements in the southern-central Alborz basalts (ZVB) indicates the enrichment of these rocks with chondrite (Thompson 1982). As seen in Fig. 3a, the relative compatibility of elements increases from left to right. Negative anomalies of Rb, Nb, Ta, Nd and Hf and positive anomalies of Zr, Th, P, La and Ti and enrichment of incompatible elements and LILE elements with chondrite are well observed in this pattern.

Figures 3a to c show the frequency of rare elements in the southern-central Alborz basalts (ZVB) relative to the early mantle (Sun & McDonough 1989, Wood *et al.*, 1979). Positive anomalies of Zr Ti, Pb, Nd, Th and enrichment of the LIL elements sometimes up to 100 times the early mantle are well observed in these diagrams. Significant negative anomalies of Hf, Ce, K, Rb and insignificant negative anomalies of Eu elements are seen in the volcanic rocks of the study area. In general, one can conclude that the level of elements in the rocks of the study area shows an enrichment between 1 and 200 time the early mantle.

Depletion of rare elements with respect to adjacent elements with negative anomalies normalized relative to a mantle or chondrite reference and remaining those elements in the residual phases at the origin and enrichment and positive anomalies are interpreted as a result of partial melting and release from mantle phases with high levels of these elements and their incompatible behavior. This is dependent on the distribution coefficients of elements with respect to the mantle phases. Therefore, contamination with crustal material and digestion phenomenon may cause positive and negative anomalies with respect to normalized references. This is not the case with basalt rocks of the region and this will be discussed below (Jung, 2003; Wilson 1989; Cox & Hawkesworth, 1985).

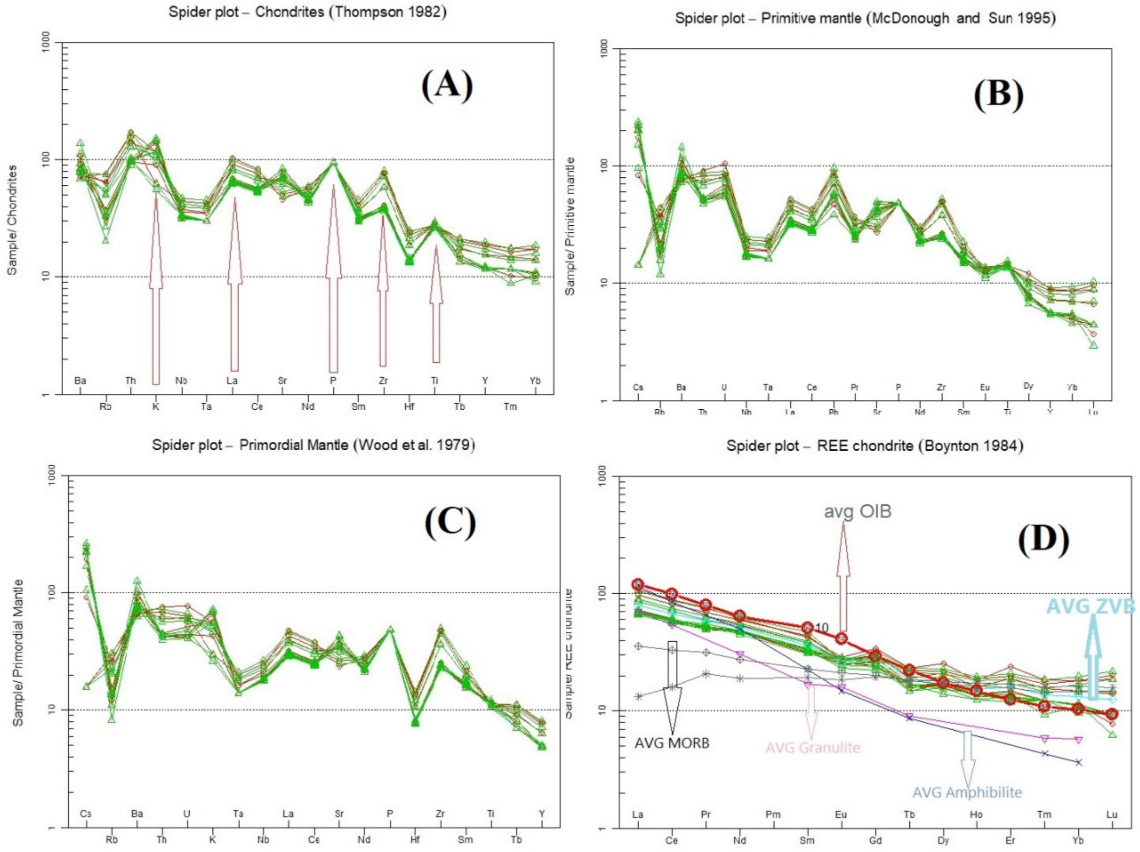


Fig. 3: The frequency of rare elements of southern-central Alborz basalts (ZVB) normalized to chondrite (Thompson, 1982) and the early mantle (Wood *et al.*, 1979, Sun & McDonough, 1989)

The negative anomaly of Nb is the characteristic of continental rocks and represents the crust involvement in magmatic processes. There are two reasons for depletion of Nb in the continental crust. First, like Ti, Nb usually remains in titanium and amphibole minerals in the subducted crust and prevents its entry into the melt and subsequent crust-formation processes (Rollinson 1993). Subduction solutions are not able to transfer Nb and thus Nb is transported by silicate melt (Pearce & Peate 1995). Although the negative anomalies of Nb and Ti are seen in the subduction regions (Ulmer 2001; Pearee 1982) and positive anomalies of these elements are specific to extensional regions (Ballever *et al.*, 2001), negative anomalies of Nb, Ti and Ta are also observed in the transitional basalts of the rift regions (Wilson 1989) and continental flood basalts (Thompson *et al.*, 1985). According to Cox & Hawkesworth (1985), the negative anomalies of Ta-Nb in the continental flood basalts (CFB) and transformation basalts of rift regions (such as the Rio Grande rift in the Taos Plateau volcanic field) are due to the presence of a residual Ta-Nb phase at the origin site or crustal contamination.

Negative anomalies of Ti can also be attributed to the presence of a Ti-containing residual phase such as amphibole, iron oxide, and titanium and clinopyroxene in the origin (Jung 2003), higher degrees of partial melting of the origin (Fodor 1987) or crustal contamination (Wilson 1989; Fodor 1987).

According to Wilson (1989) and Fodor (1987), the level of TiO_2 should be considered as a very sensitive criterion of contamination. Low concentration of TiO_2 in continental basalts is primarily related to high degree of partial melting and secondarily to crustal contamination. Among the above elements, Ti^{4+} is highly incompatible so that its concentration in chondrite is 440 ppm (McDonough & Sun, 1995), in the early mantle 1300 ppm (Sun & McDonough, 1989), in NMORB 7600 ppm (Sun & McDonough, 1989), and in EMORB is 6000 ppm (Sun & McDonough, 1989) and its average concentration in OIB is 2000 ppm (Sun, 1980). These data indicate the high incompatibility and meltophilia nature of Ti. Given the concentration of Ti in ZVB samples, the independent Ti-containing mineralization phase, i.e. titanomagnetite, is observed as the minor mineral and clinopyroxenes are observed in the titanaugite phase.

Comparing data on Afar Plume transitional basalts (Barberi *et al.*, 1975; Wilson, 1989) and flood basalts in Brazil (Thomson, 1983; Wilson, 1989) with ZVB volcanic rocks, no negative anomalies of Nb, Ti and Ta are seen in Fig. 3d. The normalized array relative to the early mantle (McDonough & Sun, 1995) and chondrite (Sun & McDonough 1989) in ZVB rocks, Afar Plume and Parana flood basalts and ocean island basalts (OIB) indicate the same behavior of elements in Ziaran volcanic rocks (ZVB). On the other hand, a negative slope and changes in the enrichment of some elements sensitive to various degrees of partial melting such as Nb are seen from left to right in the spider diagrams. This could be due to different degrees of partial melting of the parent mantle which does not confirm the role of partial crystallization and crustal contamination (Wilson 1989).

Studies show the poverty of Nb, Ti, Zr, Ce and P in rocks is similar to the pattern of rare elements in the subduction regions or magmas contaminated with crustal rocks. Given the geochemical evidence presented so far, crustal contamination or active continental active margins or back-arc basin nature of ZVB rocks is not confirmed.

As seen in Fig. 3d, given HREE enrichment in most ZVB rocks and the normalized pattern to chondrite (Boynton 1984) in ZVB rocks, LREE elements such as La show a 100-fold enrichment relative to CI while HREE elements such as Yb exhibit 20-fold enrichment with respect to CI. The $\frac{La}{Yb}$ ($_{Cn}$) ratio with an average of 5.99 is between 5.33 in the sample AB13 and 6.86 in the sample AB42. Since the rocks containing normative nepheline should exhibit a higher $\frac{La}{Yb}$ ($_{Cn}$), the highest ratio in the Ziaran volcanic rocks was observed in the AB42 sample, which is an olivine basalt. This clearly indicates heterogeneity in the early mantle or variations in the degree of partial melting of different sources of the heterogeneous mantle.

The relative concentration of LREE elements versus HREE, like $\frac{La}{Yb}$ ($_{Cn}$) ratio, shows an average of 11 in OIB rocks, and an average of 0.5 to 2 in MORB rocks.

In high ratios of LREE to HREE elements, there is a possibility of low degrees of partial melting or the absence of garnet in the rocks of origin and the lack of garnet involvement in melting as a residual phase, because of significant difference between the separation coefficient of light and heavy rare soils in the garnet. However, an increase in the degree of partial melting is probable in low ratios, but heterogeneous mantle sources and origins can play a major role in controlling the aforementioned ratios. Also, the separation of olivine and pyroxenes minerals can control LREE differentiation relative to HREE.

All samples represent an obvious negative anomaly of Eu. Therefore, in the petrographic studies of ZVB samples, a significant separation of plagioclase and the presence of plagioclase in the form of phenocrysts and matrix are visible in microscopic sections of the rocks. On the other hand, since Ca^{2+} is replaced by Eu^{+2} in the plagioclase structure, if CaO and Eu have a positive significant correlation, one can conclude that the negative anomaly of Eu is related to the presence of plagioclase. In ZVB samples, Eu concentrationis not significantly reduced with decreasing CaO. The reason for this reduction in Eu anomaly should be sought in other events.

According to Rollinson (1993) and Drake & Weill (1978), the absence of Eu anomalies and flat formation of Eu relative to the adjacent elements in the normalized patterns is due to high fugacity of oxygen and the lack of participation in the plagioclase mineral and an incompatible behavior similar to other rare earth elements. Rollinson (1993) believes that the positive anomaly of Eu is sometimes due to the presence of amphibole, clinopyroxene and garnet in parent mantle. He also believe that the enrichment of MREEs is controlled by these minerals and thus the negative anomaly of Eu can be related to the presence and participation of some of these minerals in melting conditions. Rollinson (1993) suggests the use of Eu/Eu* ratio for better understanding of Eu element. So the numbers larger than 1 show a positive anomaly and those smaller than 1 indicate negative anomaly of Eu. ZVB samples varies from 0.7 to 0.99 and this low range of changes confirms the insignificant negative anomaly of Eu in the volcanic rocks of Ziaran area and the presence of amphibole, clinopyroxene and garnet phases in melting conditions.

MREE slope relative to HREE and MREE slope relative to LREE in ZVB samples are similar to those in ocean island basalts (OIB). Comparing the ratio of LREE to MREE elements with (La/Sm $_{Cn}$), this ratio with an average of 2.2 is between 1.92 in the sample AB8 and 2.57 in the sample AB31. Comparing the ratio of MREE to HREE elements with Ce/Yb_{cn}, it can be seen that this ratio with an average of 5 is between 4.43 in the sample AB11 and 5.63 in the sample AB22. Ce/Yb_{cn} of the basalts of the Hawaiian Islands ranges from 2.09 to 3.87 (Frey *et al.*, 1991), which is close to that for ZVB rocks. Accordingly, the early magma is presumably derived from a garnet source (Kaan, 2009).

As seen in Fig. 3, the relative values of LREE to MREE elements and MREE to HREE elements are clearly close in the rare earth soils on a steady negative slope.

According to Fig. 3d, the relatively high HREE/LREE slope in the majority of samples can be an indicative of the presence of a garnet phase in the mantle source. However, it does not mean the absence of the spinel phase in the origin environment. Therefore, the produced melt belongs to a transitional zone with a spinel-garnet composition at a lithospheric pressure of 2-2.5 GPa instead of a garnet facies.

One of the important points in the geochemical interpretation of the central Alborz volcanic rocks (ZVB) and their normalized patterns is high Zr/Hf ratio with an average of 101 ranging from 91.08 in the sample AB5 to 115 in the sample AB13. According to Tappa (2004), these high Zr/Hf ratios are a reflection of a residual clinopyroxene phase. Further, Hf is 1.5-2 times more compatible than Zr in clinopyroxene mineral in the melt/ clinopyroxene system. The Zr/Hf ratio in chondrite equals 36.6 (Hofmann *et al.*, 1986) and thus one can conclude that the mantle source of the parent magma in ZVB volcanic rocks shows a depletion of Hf compared with Zr. Low degrees of partial melting are responsible for high Zr/Hf ratios .Low degrees of partial melting increase the concentration of incompatible elements such as Zr compared with compatible elements such as Y in the melt. Therefore, low degrees of partial melting cause a high Zr/Y ratio in the parent mantle of the magmatic source so that Zr/Y ratio increases with increasing Zr concentration (Pearce and Norry 1979; Pearce 1980; Kaan *et al.*, 2009).

Zr/Y ratio in the volcanic rocks of Eocene age in Ziaran area (ZVB) varies from 10.7 in the sample AB8 to 14.2 in the sample AB13. According to Kaan *et al.* (2009), these changes are due to different degrees of partial melting. They suggested that melting of an amphibole-containing source may increase the Zr/Y ratio. Given the Zr/Y ratio of 9.66 in OIB and 3.232 in E-MORB (Sun and Mac Donogh 1989), the resulting magma can be derived from an enriched mantle source of ocean island basalts (OIB) or a mantle plume at depths of more than 80 km at a pressure of 2 to 2.5 GPa participates in the production of ZVB magma.

A high Zr/Nb ratio in ZVB volcanic rocks which varies from 21.9 in the sample AB27 to 35.5 in the sample AB31 and a high Zr/Y ratio in ZVB rocks compared to Zr/Nb = 31.8 and Zr/Y = 2.64 in MORB () confirm the lack of participation of a depleted mantle in the production of the early magma of ZVB volcanic rocks and the effect of mantle contamination and the presence of a metasomatized *glimmerite* and heterogeneous mantle in the production of early magma (Kaan *et al.*, 2009).

The enrichment of the incompatible element, K, in the ZVB volcanic rocks in the normalized pattern relative to chodrite (McDonough & Sun, 1995) and the early mantle (Sun & McDonough, 1989) indicates the absence of potassium-containing residual phases during the melting of the mantle source (Wilson and Downes, 1991,), but one cannot clearly conclude that K exists is in phlogopite, amphibole, clinopyroxene or garnet phase.

Potassium is depleted from a potassium-containing phase during partial melting of the mantle source and its concentration is affected by different degrees of partial melting. Therefore, K/Nb and K/La ratios can be a good indicator in determining the degree of partial melting and heterogeneous mantle sources.

Amphiboles have a high separation factor of HFSE elements and show higher concentrations of HFSE elements compared to the potassium-containing phlogopite phase in the melt. Rb in the potassium-containing amphibole phase shows a compatible behavior relative to phlogopite. If it originates from a phlogopite source, it will show an enrichment when it is normalized to chondrite and the early mantle (Foley *et al.*, 1996; Class and Goldstain 1997).

Due to the significant depletion of Rb and the enrichment of HFSE elements in ZVB rocks, an amphibole mantle is not far from the mind as the source of early magma. It seems unlikely that phlogopite could be a K-containing phase in the mantle source.

The spinel mineral is replaced by the titanite mineral in the peridotite mantle during the metasomatism process and contains HFSE elements such as Nb and Ta which is known as a minor mineral in the mantle source. But as mentioned, separation of these elements can be due to participation of the amphibole mineral or the absence of titanite in melting conditions.

Pb is another mobile element showing a significant enrichment in the ZVB specimens relative to the early mantle and OIB. Such a pattern of Pb enrichment is not observed in standard samples. Pb is a completely volatile element (White, 2005) with chalcophile and siderophile properties. Pb has a capacity of 2+ in a wide range of oxidation and reduction conditions. This element with an electronegativity of 1.9 implies a high degree of covalent bonding. Although it shows a low dissolution in most natural conditions, Pb is easily dissolved by forming a complex with elements such as chlorine Cl in hydrothermal and metamorphic solutions (White, 2005). Pb may be formed due to crustal contamination or hydrothermal metasomatism processes. Positive anomalies of Pb can also be caused by magma contamination with crustal material due to the concentration of this element in the continental crust. Unlike Nb, Pb is gradually enriched in the crust by transferring from subduction solutions of subduction systems and crust-forming processes and produced melts (Jung 2003).

The negative anomaly of Ce is mainly observed in seawater, pelagic sediments and radiolarite cherts. The positive anomaly of Ce has been reported in the Fe-Mn nodules ; Therefore, this negative anomaly can be due to the depletion of the origin rock from this element or due to oxidation conditions and hydrothermal alterations and solutions. But one of the most important reason for negative anomaly of Ce in volcanic rocks may be contamination and digestion of the crustal material with magma in subduction zones confirming the role of subduction sediments along with subducted oceanic crust (Slab) in continental margins. Since this magmatism process occurred during the Eocene period in Alborz and this region was not an active continental margin at that time, this level of Ce depletion can be attributed to the mantle contamination by old subducted oceanic crusts or contamination by a continental lithosphere falling in the mantle or mechanical erosion of the continental lithosphere and metasomatism of the mantle (Jakes and Gill, 1970).

Weathering process is one of the important factors in the negative anomaly of Ce in old basalts and volcanic rocks in tropical regions and even in intercontinental basalts. This phenomenon causes an unusual increase in the concentration of REE elements compared to LILE elements such as (Rb, K, Sr) and HFSE. According to the variations of Nb, Y, and La against each other, the role of weathering in the negative anomaly of Ce in the Ziaran volcanic rocks looks far from mind, because Y is more mobile than Nb and La and goes out of the rock during weathering process. The variations of the elements against each other (Figs.) suggesting the ineffectiveness of the weathering process and its relationship with various degrees of partial melting of magmatic sources. The changes in the concentration of REE elements confirm this conclusion (Class and Le Roex, 2008). Negative anomaly of Ce is obtained while contamination with rocks and deposits near the surface of the earth causes an increase in the concentration of Pb and a significant decrease in Pb/Ce ratio (about 1-5) (Class and Le Roex, 2008). The average value of Pb/Ce ratio in OIB was 25 ± 5 (Sun & McDonough, 1989). The negative anomaly of Ce in ocean island basalts (OIB) is due to cycle of old sedimentary deposits in mantle sources .

The average Pb/Ce ratio of ZVB volcanic rocks is 0.17 and varies from 0.16 to 0.2 and thus is not within the range of OIB variations.

Therefore, this should be studied in detail like the absence of Nb and Ta in the melt conditions to understand whether Ce is located in residual phases and its concentration has been decreased in the melt or the melt underwent crustal contamination or old tectonic activities have caused the mantle under the crust of Alborz to be contaminated.

A small negative anomaly of Y is seen in the normalized pattern of rocks and the average value of ZVB volcanic rocks in Fig. 4-10. This is also observed in the normalized pattern of the mean OIB relative to the early mantle. On the other hand, since Y is considered to be less incompatible and a heavier element and Y-containing garnet is compatible, the role and presence of garnet in the parent magma source of ZVB is fully confirmed.

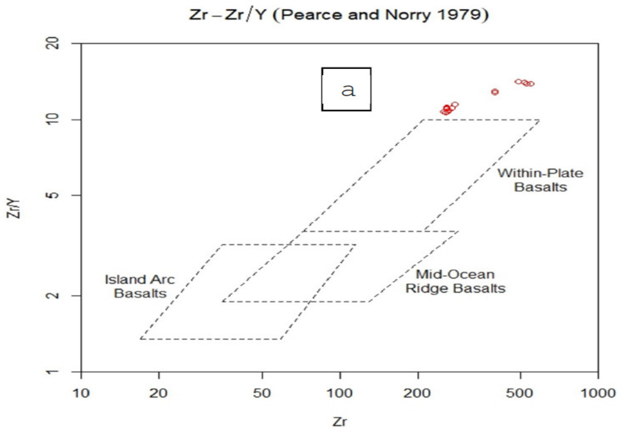
The ratios of incompatible rare elements and the patterns of rare elements with respect to normalized references show that magma formed ZVB basalts is not contaminated with the continental crust. Given the ratios of the rare elements, as well as the patterns of these elements in the spider diagrams, it is clear that these magmas are mostly originated from deep areas, enriched mantles along with garnet and amphibole phases with different degrees of partial melting.

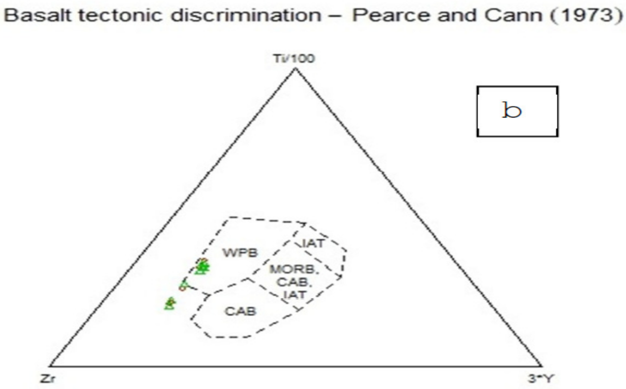
Tectonomagmatic environment

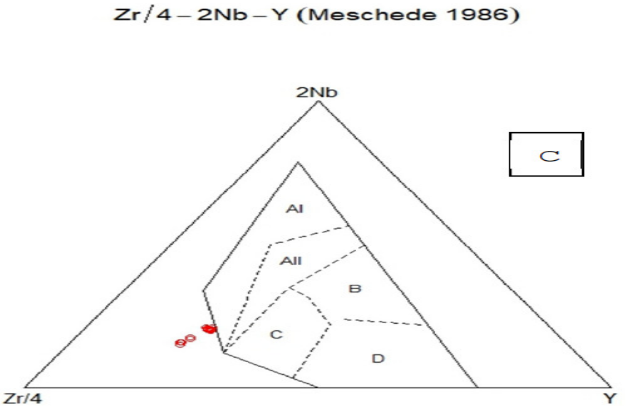
To determine the tectonomagmatic environment of ZVB rocks, the diagrams provided by Sun & McDonough (1989), Wood (1980, 1989) and Meshede (1989) (Fig. 4) were used. In these diagrams, non-mobile incompatible elements including Ti, Y, Th, Nb, Ta and Zr are used to divide different fields. All ZVB samples are in the range of within-plate basalts (WPB) and mainly show a mantle metasomatism process with different degrees of partial melting with Zr concentration.

As seen in the Zr / Y diagram versus Zr (Pearce & Norry, 1979) (Fig. 4a), the tectonic setting of ZVB lies within the range of within-plate basalts and some samples show the enrichment of Zr which is mainly related to mantle metasomatism by melts of slow degrees of partial melting at very low depths.

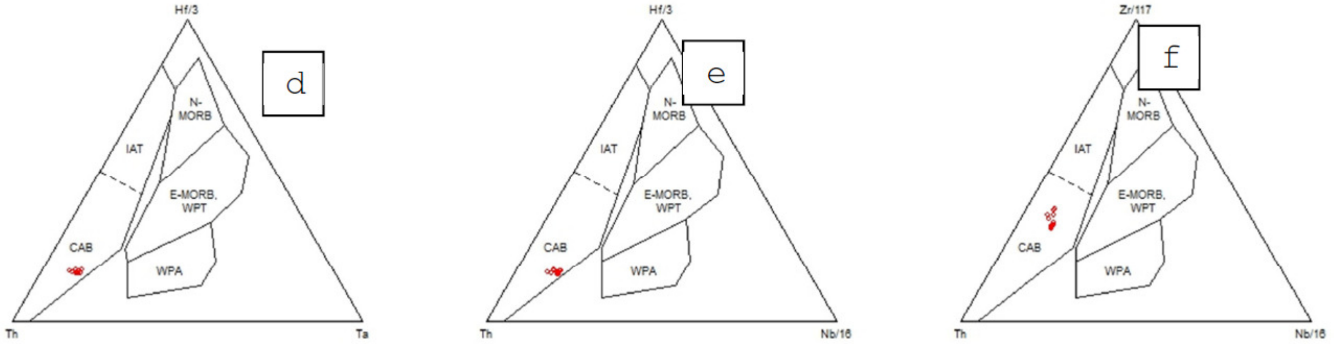
As seen in the triangular diagram of Ti/100-Zr-Y*3 in Fig. 4-11b, most samples are within the range of D within-plate basalts. In the triangular diagram of Nb*2-Zr/4-Y (Meschede, 1989), most samples are in the range of AI within-plate alkaline basalts and even more alkaline samples with lower degrees of partial melting and higher Zr levels, as within-plate basalts, are in the same range (Fig. 4-12c).


Zr – Zr/Y (Pearce and Norry 1979)


Basalt tectonic discrimination – Pearce and Cann (1973)


Zr/4 – 2Nb – Y (Meschede 1986)

Triangular diagrams of the Th-Hf-Ta-Zr-Nb system, Wood 1980



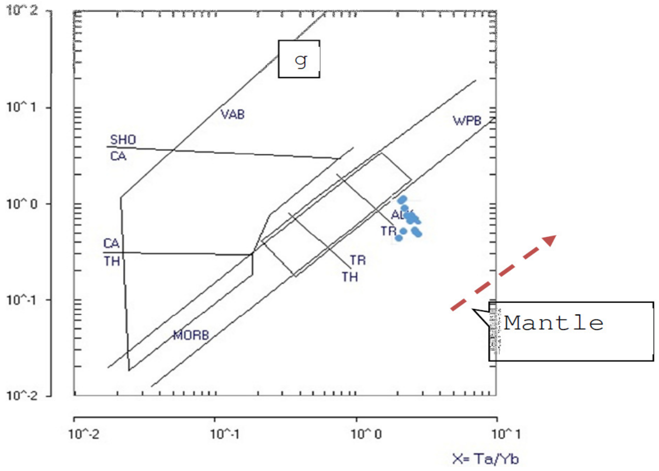


Fig. 4: Determining the Tectonomagmatic environment of volcanic rocks of the upper middle Eocene age in Ziaran area (ZVB) using Sun & McDonough (1989), Wood (1980, 1989) and Meshede (1989) diagrams

As seen in the triangular diagrams of Hf/3-Th-Ta, Hf/3-Th-Nb/16 and Zr/118-Th-Nb/18 (Wood 1989) (Figs. 4e, f and d), all samples are located in the range of active continental margin arc basalts (CAB) while the sample should be in the range of within-plate basalts (WPBs). According to aforementioned discussion and positive correlations between the constituent elements and the same origin of elements, mantle metasomatism by old continental crusts increased concentration of Th in magmatic sources. The compatible behavior of Ta in the residual mineral phase reduced the proportion of these elements in the produced magma. Therefore, the samples are located in the range of active continental margin arc basalts (CAB).

In the T/Yb-Th/Yb diagram (Fig. 4g), the ratio of these elements is strongly sensitive to the mantle composition, pressure variations and varying degrees of partial melting. This confirms the role of ocean island basalts (OIB) source/sources, enrichment in *asthenosphere* mantle and different degrees of partial melting in the formation of the early magma of Eocene volcanic rocks in the southern Alborz in Ziaran and south of Taleqan. In terms of chemical composition, the rocks are located within the range of alkaline to transitional basalts.

According to studies on the ratios of rare elements and patterns of incompatible elements and comparing them with crustal concentrations and rift zones, southern-central Alborz basalts (ZVB) are in connection with the magmatism of intercontinental rift zones. In the upper middle Eocene, the southern Alborz was dominated by a shallow sea. *Asthenosphere* uplifted in the area with outcrops of ZVB volcanism. This *asthenosphere* uplift was associated with a decrease in the depth of the sedimentary basin and a change in the sedimentation regime. *Asthenosphere* uplift and local fractures due to extensional (rifting) phases caused magma formation in depth and linear volcanism of ZVB volcanic rocks along these fractures. As seen in Fig. 4, volcanism of ZVB volcanic rocks is not related to divergent or convergent boundaries of lithospheric sheets. From the geochemical point of view, rare and rare earth elements of ZVB samples displayed evidence of OIB sources so that incompatible elements showed a significant enrichment which is not comparable to MORB rocks.

On the other hand, the absence of a prominent negative anomaly of Nb as well as other HFS elements such as Hf, Zr, Ta and Ti indicates that ZVB basalts are not of island arc basalt (IAB) type despite enrichment of water-soluble elements, such as Ba and Pb in these rocks. Comparing the diagram in Fig. 3d on the average concentration of rare elements in a normalized pattern relative to the early mantle, IAB samples show a lower enrichment of LREE and MREE elements than ZVB samples. Also, the enrichment of HFS and LFS elements is different from those of ZVB. In particular, the highly negative anomaly of Nb in the IAB pattern is quite clear and different from ZVB samples. Therefore, the relationship between ZVB volcanism and oceanic crust subducting mechanism underneath the continental crust is also quite different. The pattern of incompatible elements with MORB indicates the enrichment of HFS and LFS incompatible elements as well as the LREE and MREE elements of the ZVB samples compared with MORB. On the other hand, ZVB samples are mostly in the range of alkali basalts (Fig. 3).

Considering the above discussion and given that the southern-central Alborz basalts (ZVB) show no strong evidence of crustal contamination based on geochemical studies, but were approximately comparable with classic intercontinental alkaline basalts, the characteristics of the source region of these basalts with a variety of mantle sources are fitted with the ratio of incompatible elements.

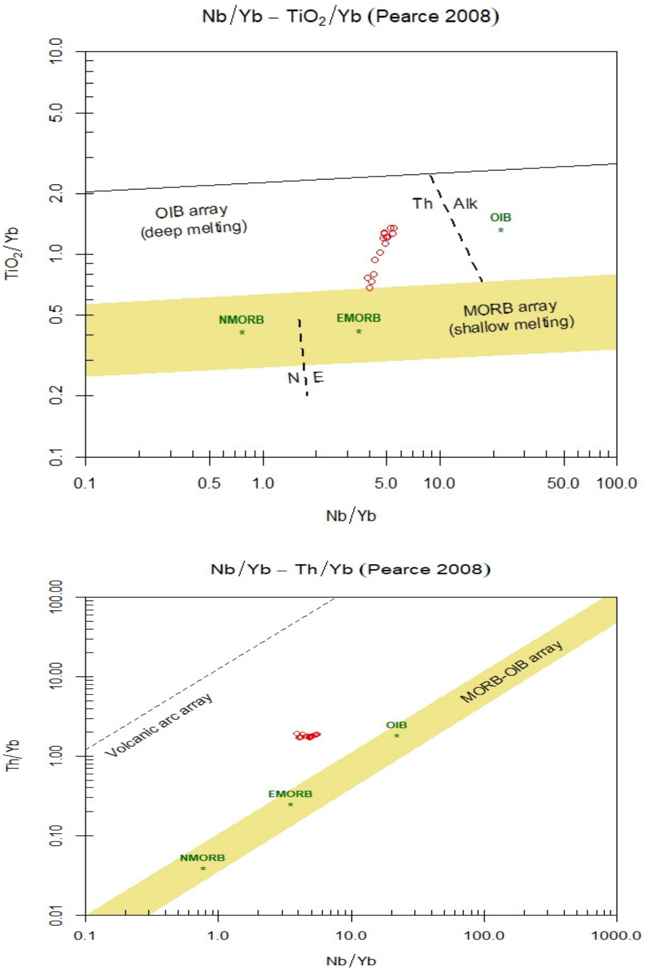


Fig. 6: Determining the Tectonomagmatic environment of the upper middle Eocene volcanic rocks of Albroz in Ziaran (ZVB) using Pearee (2008) diagrams

In the Nb/Yb diagram versus TiO_2/Yb, the mean values of N-MORB, P-MORB and OIB (Sun & McDonough 1989) are given to be compared with alkaline basalts of Ziaran (ZVB). As seen, most samples show very deep OIB sources and the lack of contamination with the middle and upper crust and even lack of connection with MORB sources. This is fully confirmed in the Nb/Yb versus Th/Yb diagram showing the lack of relationship with crustal sources (Figs. 6-a, b).

Conclusion

Basic to intermediate volcanic rocks of upper middle Eocene alkaline chemical composition in Ziaran region in the central part of Alborz Mountains showed a lithological composition ranged from alkaline-olivine basalt to andesite. The normalized pattern of the frequency of incompatible rare elements in Ziaran volcanic rocks indicated the enrichment of these rocks with chondrite in the early mantle. Negative anomalies, remaining those elements in the residual phases at origin and positive anomalies were interpreted as a result of partial melting and release of mantle phases with high contents of these elements and their incompatible behavior. Therefore, contamination with crustal material and digestion phenomenon may cause positive and negative anomalies with respect to the normalized reference. The relative behavior of the ratio of LILE, HFSE and REE elements was studied to understand the behavior and sources of the parent magma of the upper middle Eocene volcanic rocks in the region. The results showed an enriched source of oceanic island basalts (OIB). Therefore, the role of the upper, middle, lower and lower continental lithosphere in contamination of the early magma is insignificant. This confirms the presence of mantle amphiboles and the absence of garnet and clinopyroxene in melting conditions.

References

Allen, M.; Ghassemi, M.R.; Shahrabi, M.; Qorashi, M. (2003). Accommodation of late Cenozoic oblique shortening in the Alborz range, northern Iran. Journal of Structural Geology 25: 659–672.

Ballever, M.; LeGoff, E. and Hebet, R. (2001). The tectothermal evolution of the Caledonian belt of Northern Brittany, France: A Neoproterozoic volcanic arc.Tectonophysics. 33: 119-43.

Berberian, M. (1983). The southern Caspian: a compressional depression floored by a trapped, modified oceanic crust. Canadian Journal of Earth Sciences 20, 163–183.

Boynton, W.V. (1984). Geochemistry of the rare earth elements: meteorite studies. In: Henderson P (ed) Rare earth element geochemistry. Elsevier, Amsterdam, pp 63–114

Coogan, L.A.; Thompson, G.; MacLeod, C.J.; (2001). A textural and geochemical investigation of high level gabbros ftom the oman ophiolite, Littos 63: 67-82.

Cox, K.G.; Bell, J.D. and Pankhurst, R.J. (1979). "The interpretation of igneous rocks". George Allen and unwin, London, 450P

Cox, K.G.; Bell, J.D. and Pankhurst, R.J. (1979). "The interpretation of igneous rocks". George Allen and unwin, London, 450P.

Fodor, R.V. (1987). Low and high – TiO_2 flood basalts of southern Brazil.Origin from picritic Parentage and a common mantle source. Earth planet. Sci. Lett. 84: 423-430.

Green, D.H. and Ringwood, A.E. (1967). The genesis of basaltic magmas. Contrib. Mineral. Petrol. 15: 103-190.

Guest, B.; Guest, A.; Axen, F. (2007). Late Tertiary tectonic evolution of northern Iran: a case for simple crustal folding. Global and Planetary Change, 58: 435–453.

Hofmann, A.W.; Jochum, K.P.; Seufert, M. and White, W.M. (1986). Nd and Pb in Oceanic basalts, New constraints on mantle evolution Earth Planet. Sco. Lett., 90: 297-317.

Jakes, P. and White, A.J.R. (1970). K/Rb ratios of rocks from island arcs. Geochem. Cismoshim Acta. 34: 849-856.

Jenner, G.A.; Foley, S.F.; Jackson, S.E.; Green, T.H.; Fryer, B.J. and Longerich, H.P. (1994). Determination of partition coefficients for trace elements in high pressure–temperature experimental run products by laser ablation microprobe-inductively coupled plasmamass spectrometry (LAM-ICP-MS).Geochim Cosmochim Acta, 58: 5099–5130.

Jung, C. (2003). "Geochemische and Isotopen-geochemische Untersuchungen an tertiaeren Vulkaniten der Hocheifel-ein beitrag zur identifizierung der mantelquellen von Rift-bezogenen Vulkaniten”, Dissertation zut Erlangung des doktorgrades Naturwissenchaften,vorgelegt dem fachbreich Geowissenschaften der Philipps-Universitaet Marburg.

Kaan, S.M. and Cemal, G. (2009). Geochemistry of mafic rocks of the Karakaya complex, Turkey: evidence for plume-involvement in the Palaeotethyan extensional regime during the Middle and Late Triassic. Int J Earth Sci., 98: 367- 385.

Le Bas, M.J.; Le Maitre, R.W.; Streckeisen, A. and Zanettin, B.A. (1986). Chemical classification of volcanic rocks based on the total alkali-silica diagram. J Petrol, 27: 745–750.

Lemaitre, R.W.; Bateman, P.; Dudek, A.; Keller, K.; Lameyre Lebas, M.J.; Sabine, P.A.; Schmid, K.; Sorensenlt, A.; Wooley, A.R. and Zanettin, B. (1989). A classification of ignueos roks and glossary terms, Blackwell, Oxford.

Mahoney, J.J.; Natland, J.H.; White, W.M. *et al.* (1989). Isotopic and geochemical provinces of western Indian Ocean spreading centers. J Geophys Res, 94: 4033-4052.

McDonough, W.F. (1990). Constraints on the composition of the continental lithospheric mantle. Earth and Planetary Science Letters 101: 1–18.

Meschede, M. (1986). A method of discriminating between different type of mid-ocean ridge basalts and continental tholeiites with the Nb–Zr–Y diagram. Chemical Geology, 56: 207–218.

Rickwood, P.C. (1989). "Boundary lines within petrologic digrams which use oxide of major and minor elements”. Lithos. 22: 225-240.

Rollinson, H.R. (1993). Using Geochemical Data: Evaluation, Presentation, Interpretation. Longman, New York, p.352.

Stocklin, J. (1974). Northern Iran: Alborz mountains. In: Spencer, A. (Ed.), Mesozoic–Cenozoic orogenic belts: data for orogenic studies. Geologicalm Society Special Publication 4: 213–234.

Stocklin, J. (1974). Possible ancient continental margins in Iran. In: Burk, C.A.; Drake, C.L. (Eds.), The Geology of Continental Margins. Springer, Berlin, pp.873e887.

Stocklin, J. (1974b). Possible ancient continental margins in Iran, *in* Burke, C.; and Drake, C.;eds.; The geology of continental margins: New York, Springer-Verlag, 873–887.

Sun, S.S. and McDonough, W.F. (1989). Chemical and isotopic systematics of oceanic basalts: implications for mantle composition and processes. In: Saunders AD, Norry MJ (eds) Magmatism in the ocean basins.Geol Soc Spec Publ 42: 313–345.

Ulmer, P. (2001). Partial melting in the mantle wedge-the role of H_2O in the genesis of mantle-derived" arc-related" magmas. Physics of the Earth and Planetary Interiors, 127: 215-232.

Whitmarsh, R.B.; Minshull, T.A.; Russel, S.M. (2001). The role of syn-rift magmatism in the Iberia margin: geophysical observations. In: Wilson, R.C.L.; Whitmarsh, R.B.; et al, (eds) Non-volcanic rifting of continental margin: a comparison of evidence from land and sea. Geol. Soc. Lond. Spec. Publ. 187: 07-124.

Winchester, J.A. and Floyd, P.A. (1977). Geochemical discrimination of different magma series and their differentiation products using immobile elements. Chemical Geology 20: 325–342.

Wood, D.A.; Joron, J.L.; Treuil, M.; Norry, M. and Tarney, J. (1979). Elemental and Sr isotope variations in basic lavas from Iceland and the surrounding ocean floor. Contrib. Mineral. Petrol., 70: 319-339.