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Petrological and Geodynamical Constraints of Chaldoran Basaltic Rocks, NW Iran: Evidence from Geochemical Characteristics

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Abstract

The Chaldoran area in the northwestern portion of Iran has a Mesozoic oceanic crust basement. The studied rocks of this region can be divided into three groups: ophiolitic gabbros and pillow lavas, ophiolitic volcanoclastics and Eocene lava flows. Ophiolitic mafic rocks exhibit a continental volcanic arc nature while Eocene lava flows are of an OIB-like nature. During the Mesozoic era, the Chaldoran region was situated in the active continental margin of the Iranian plate due to Neotetyhan subduction beneath the SSZ (Sanandaj-Sirjan Zone). As the Neo-Tethyan subduction regime changed, a rift in the Chaldoran area was generated which caused a continental back arc basin during the Upper Cretaceous period. A spreading basin in the oceanic ridge along with the creation of newly formed oceanic crust gave rise to oceanic islands which became active and formed volcanoclastic rocks in the Chaldoran Oceanic crust. These were the result of ascending continental arc nature plumes that formed in the sub-oceanic crust and extended up to the Neo-Tethys subducted slab in the mantle wedge. After the basin finished spreading in the early Paleocene, sedimentation followed continuing up to the Middle Eocene. During the Upper Eocene, alkaline and OIB-like volcanism was active in the Chaldoran area due to the breaking off of the Neo-Tethyan subducted slab in the sub-oceanic crust of the Chaldoran area causing the development of an asthenospheric window.

Keywords: *Iran, Chaldoran, back arc basin, Neo-Tethys, OIB-like volcanism.*

Introduction

The Iranian plateau is a tectonically active region within the Alpine–Himalayan orogenic belt. It contains a number of continental fragments that have been welded together along a suture zone of oceanic nature [1]. Iran began mainly as a result of the closure of the Neo-Tethyan Ocean during Cenozoic times. However, the neotectonic episodes of Iran were shaped in the Neogene during interaction between the northward-moving Arabian plates and the relatively stable Eurasian plate. Subduction of the Neo-Tethys Ocean beneath the Eurasian lands formed a complex set of subduction zones from Greece to Iran with several volcanic provinces of different ages and composition.

Different models have been proposed for the geodynamics of Northwestern Iran, but the most comprehensive model is offered by [2]. According to this model, Eocene volcanic rocks in the northern Tabriz fault which show characteristics of active continental margin rock affinity, are related to the Khoy-Zanjan oceanic subduction beneath the Azarbaijan plate. It is believed that the ultramafic rocks along the Tabriz fault, Khoy and Chaldoran (Siyah-Cheshmeh) ophiolites, are the relics of this oceanic crust. According to the new model [3] the basement rock in northwestern Iran is related to the Cretaceous oceanic crust. The Eocene and Oligocene magmatism in northwestern Iran is the result of the closure of the Neo-Tethys between the Arabian plate and the Sanandaj-Sirjan zone. It is suggested that the alkaline and subduction-related natures of the young volcanic rocks (Miocene to Quaternary) were formed along the strike-slip faults which are a result of the partial melting of the up-welling asthenosphere at the hinterland of the collisional zone between Arabia and Eurasia [4]. Iranian ophiolites are part of the Tethyan ophiolite belt of the Middle East which links the eastern Mediterranean (Hellenides–Dinarides) ophiolites in the west, with the Asian ophiolites in the east. The Chaldoran region is located in the northwestern corner of the Iranian Azerbaijan province and extends around the Turkish border. This area has an oceanic crust of basement rock. The geology of the Chaldoran area is still widely unknown. There are few studies about its geological features. Some of them show that the ophiolites of this area are related to the Khoy ophiolite (e.g. [5]; [6]).On the basis of studies [2] and [1], the Khoy ophiolite may be a part of the Northern Anatolian Fault (NAF) of dismembered ophiolites. It is also suggested that there is a new

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branch of Tethyan oceanic crust, called the Khoy-Zanjan branch, which trends NW–SE parallel to the Zagros fault. In this regard, [1] proposed an accordion tectonic model for this part of Iran. Khalatbari-Jaffari( )[6,7,8] introduced two types of ophiolites in the Khoy region; the eastern ophiolite which is mainly metamorphosed (upper Triassic to upper Cretaceous) and shows a subduction zone affinity, and the western portion which is not metamorphosed (Cretaceous age) and shows a T-MORB affinity. However, Azizi et al. (2011)[9] based on U-Pb age dating, showed that the metamorphic belt classified as Jurassic ophiolite, was generated during the Precambrian. The first main study of the Chaldoran region was carried out by the Geological Survey of Iran with the aim of creating a 1/250000 scale geological map of Maku. Recently, Rahghoshay et al. (2008) [5] and Poormohsen et al. (2010)[6] studied the magmatic rocks of the Chaldoran ophiolite and recognized two different types of basaltic rocks: one group with supra subduction zone affinity and another group with OIB characteristics. Because of the similarities between the Khoy and Chaldoran ophiolites and the lack of published data about the Chaldoran ophiolite, this study was based on the Khoy ophiolite. According to the field study and information pertaining to the petrographic and geochemical characteristics of ophiolite-related rocks and magmatic rocks unrelated to ophiolites, efforts were made to determine the tectonomagmatic and geodynamic evolution of the Chaldoran area during the Mesozoic and Cenozoic eras.

Geological setting

In the context of the structural subdivisions of Iran, various tectonic zones were proposed for the Chaldoran (Siyah-Cheshmeh) area including; the Central Iran Zone by Stocklin (1968)[10], the Soltanieye-Misho Zone by Eftekhar Nejad (1980)[11], and the Sanandaj-Sirjan Zone by Aghanabati (1998)[12].The study area is located in the Cretaceous–Cenozoic accreted units with ophiolite areas containing oceanic crust (Fig. 1)[13]. Geologically, the Chaldoran ophiolite is part of the Khoy ophiolite and is related to the colored mélange (Fig. 2). Northwestern Iran contains the largest quantity of remnants of the Mesozoic oceanic crust complex. The oldest formation in this region is related to the upper Cretaceous formation (Alavi et al. 1975; Rahgoshay et al. 2008; Poormohsen et al. 2010)[14, 4, 5] which includes ultramafic and mafic rocks (diabasic dykes, gabbro and pillow lava). Tertiary units in this zone include volcanic rocks and sediments which are mostly composed of limestone flysch and molasse which overlie older units unconformably. There is no chronological data on the ophiolitic rocks.

There is extensive exposure of ultramafic rocks in the western and northern parts of the Kelisa Kandi village. These groups include dark-grey to black serpentinite with pyroxenite and peridotite. The ultramafic rocks in this region are serpentinized, therefore creating a separation of the ultramafic rocks with the other rocks in the color mélange; the placement of these most likely occurred during the upper Cretaceous (Alavi et al. 1975)[14]. The lack of metamorphism in the ultramafic rocks and sedimentary rocks indicates an intrusion and uprising of ultramafic rocks which occurred in a cool state (Alavi et al. 1975)[14]. In this complex, there is no evidence of a transitional zone, chromitite lens or sheeted dyke. Some diabasic isolated dykes and gabbroic pockets which occur locally within the high-level ultramafic rocks show no chilled margins.

Pillow lava is rarely exposed around the Badavli and Mazraeh villages. The samples around the Mazraeh village have vesicular rims filled with calcite and are spilitized. In some places various dykes with different compositions are hosted by pillow lava. This unit rests tectonically on ultramafic rocks or Cretaceous sediments. In some places gabbro residue can be seen at the base of the faulted block on the ultramafic rocks or Cretaceous sediment. In most parts of the outcrops, pillow lavas are overlain by Maastrichtian marine sediments with normal contact. The thickness of the marine sediments over the pillow lava varies and in some places reaches up to 1 m.

The thin Maastrichtian limestone is dark to light and its fractures are filled with nodules of chert. The color of the limestone varies from red to pink to green in some areas. This unit is located in the northern and western portoins of Khukhur Kaandi village, the northern and southern portions of Khangol village, and the eastern portion of the road that is situated between the Chaldoran and Kelisa Kandi village. This unit is composed of cryptocrystalline quartz along with a vein of quartz. The age of this unit is most likely the same as the Maastrichtian limestone (Alavi et al. 1975)[14] and overlies the pillow lava with normal contact.

The composition of the volcanoclasts is basaltic, andesitic, dacitic and rhyolitic. Throughout the Mazraeh village, basaltic and andesitic volcanoclasts and old breccias overlay pillow lava. The size of these clasts is between 20 and 30 cm and in some instance reaches up to 50 cm. Some basaltic clasts with pillow structures are distributed in the background of pyroclastic rocks. There are layers of lapilli tuffs and crystal lithic tuffs with a thickness of up to 1 meter, inter-bedded with volcanic breccias. In some outcrops, these clasts and breccias are overlain with Maastrichtian marine sediment with normal contact and are covered in some places with Paleocene-Eocene sediments. The lowest part of the Tertiary unit is composed of a conglomeratic layer with good bedding and crushed fragments of ultramafic rocks and color mélange covered with red and pink limestone. The existence of ultramafic rocks in this conglomeratic unit indicates that emplacement of these rocks occurred before the sedimentation of flysch-type units (Alavi et al. 1975)[14]. The Eocene flysch-type unit is comprised mainly of sandstone, limestone, silt and red marl. Green to milky limestone of Paleocene-Eocene age is exposed in the western portion of the Nabikandi, Jongtape and Arabdizaj villages. This unit is located on the Maastrichtian sedimentary units or mixed with the ultramafic and conglomeratic units. The limestone was deposited after the emplacement of ultramafic rocks, however, due to tectonic activity, this unit mixed with other units. On the basis of fossil content, this unit is believed to be Paleocene-Eocene in age (Alavi et al. 1975)[14]. The ophiolitic rocks and Paleocene-Eocene units are separated by later diabasic –gabbroic dykes. These dykes are related to the magmatic events in the Chaldoran region. They show symmetrical and asymmetrical chilled margins with widths varying from 0.5 m to 5 m. Lava made up of a basaltic composition is located above the middle Eocene age sedimentary units. The thicknesses of these lavas are different. These lavas are only exposed locally and cannot be seen in the entire study area. They are covered by sediments of Oligo-Miocene age with the most likely age being upper Eocene to Oligocene. Tertiary sediments in the Chaldoran region are mainly composed of Miocene sediment (lower red and Qom formation). The Miocene units include red conglomerate, pink and yellow to white limestone in addition to green marble. They uncomformably cover the metamorphic rock, Paleozoic rocks and other old rocks. The thickness of these rocks is up to 600 meters which is divided into three units; basal conglomerate, limestone, and marble (Alavi et al. 1975) [14].

The youngest magmatic activity in this region is the Ararat basaltic flow of Quaternary age. Outcrops of the youngest basaltic rocks are exposed in the northern and northwestern parts of Chaldoran in the vicinity of the Turkish border as well as the eastern portion of Tazekand village and the northern portion of Zaviye Bala village. These basaltic lavas are most likely related to Ararat volcanismic activity in the early Quaternary period.

Petrography

Gabbros are rarely exposed in the region; the texture of these rocks is granular. Fine grains of plagioclase and pyroxene are the main constituents of these rocks. Olivine, amphibole and opaque minerals are in minor phases. Plagioclase crystals are subhedral to anhedral and are completely decomposed to clay minerals and epidote (Fig.3a). Clinopyroxenes were observed to have been altered to uralite and chlorite. Amphiboles along with opaque minerals were seen in the vicinity of altered clinopyroxenes. Pillow lavas petrographically are diverse and have a basaltic composition with an intersertal texture. Plagioclase and clinopyroxene (zoned) are the main former minerals. Plagioclase occurred either as a fresh crystal or altered microlite. Other minerals such as calcite and iron oxide are present in these rocks. The glassy mesostase in these rocks are completely altered to clay minerals. An additional type of pillow lava seen around the Kord Kandi and Mazrae villages has a basaltic composition and is characterized by an intersertal and glomoroporphyritic texture with mesostase composed of plagioclase and clinopyroxene microlites. A portion of the samples taken of these rocks is mainly porphyritic. The main constituent minerals of these rocks are plagioclase, clinopyroxene and olivine. Plagioclase occurred either as a glomoroporphyritic aggregate or as a fine grained microlite in the mesostase that was altered to sericite. Pyroxene in addition to iddengisited olivine can be seen intergranularly between the plagioclase minerals (Fig. 3b). Pyrite and magnetite are rarely seen in these rocks. Mafic fragments are petrographically basalts and show intersertal and pilotaxitic textures. Plagioclase, clinopyroxene and amphibole are the main constituent minerals. Plagioclase microlites occurred between the fine grained crystals of clinopyroxene (Fig. 3c). Brown amphiboles are arranged around the clinopyroxene. Glassy material, apatite, calcite, chlorite and polycrystalline quartz are located between clinopyroxene and plagioclase. Clinopyroxene in these rocks is coarse grained with a high relief and birefregence (Fig. 3d). The glassy groundmass is altered to clay minerals. In the andesitic fragment, plagioclase occurred either as a euhedral phenocryst or as a fine grained microlite in the groundmass and mainly altered to clay minerals.

Eocene lava flows are aphyric, although lava flows with large plagioclase phenocrysts exist between the volcanicalstic rocks. Petrographically, these rocks are different. The first group of lava flows sampled showed amphibole phenocrysts and lesser plagioclase. The texture of these rocks is intergranular, intersertal and phorphyric with a microcrystalline groundmass. Amphibole is the dominant mineral in these rocks. Clinopyroxene occurred as a large phenocryst (1mm) and also as a small crystal between the plagioclase and shows zoning. Dark to pale brown prismatic amphibole are distributed as a tiny component in these rocks as a phenocryst. They are partially altered to chlorite and plagioclase but completely altered to clay minerals and calcite (Fig. 3e). Glassy material and magnetite with an intersertal texture occur between the clinopyroxene and plagioclase. Amphibole phenocrysts are present along with plagioclase and clinopyroxene in the groundmass composed of clinopyroxene, plagioclase, amphibole, magnetite and glassy materials indicating that the composition of these rocks is alkaline (Fig. 3f). Petrographic descriptions in the studied samples suggest that the rocks under investigation suffered significant low-T alteration.

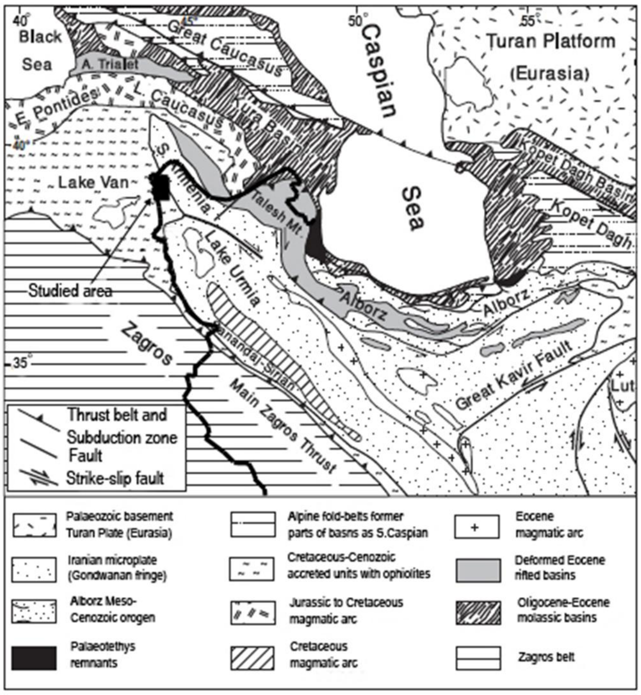

Fig. 1. Geological map of NW Iran and the adjacent areas from Moazzen and Oberhänsli (2008) [13]. The studied area is located between the Caucasus suture to the north and the Zagros suture to the south.

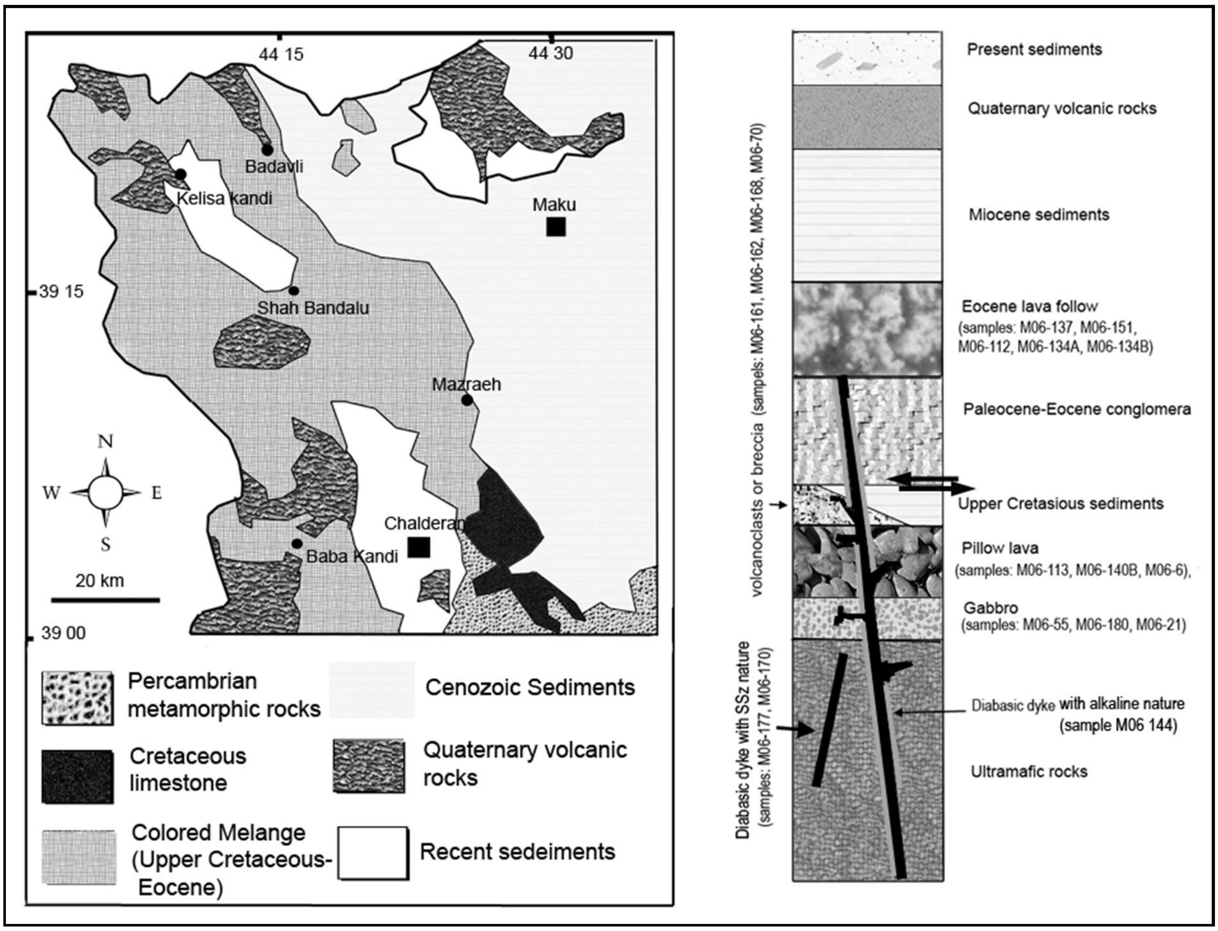

Fig. 2. Simplified geological map and pseudo-stratigraphic column of the studied area.

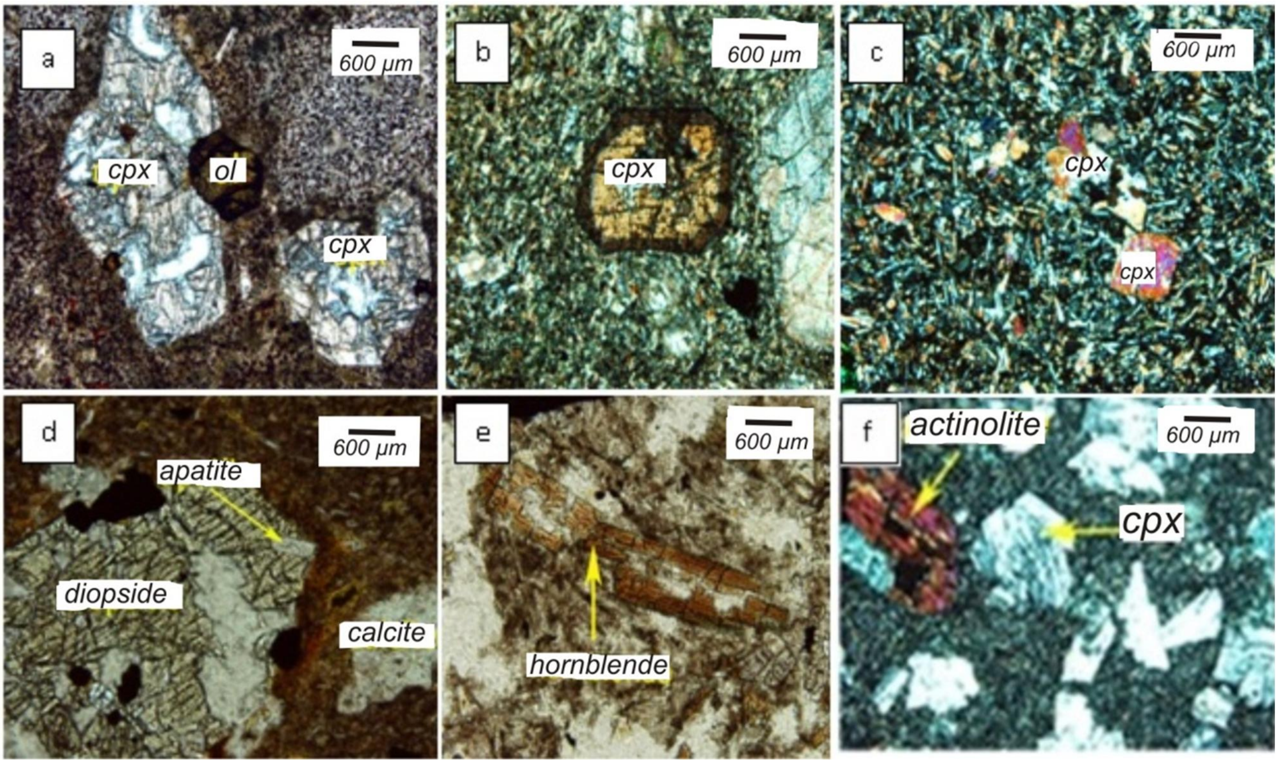

Fig. 3. Microscopic photos of some studied rocks. (a) gabbro with plagioclase completely decomposed to clay minerals, (b) iddingsitized clinopyroxene in pillow lava, (c) plagioclase microlites between clinopyroxene grain in volcanoclastic, (d) coarse grain of clinopyroxene in volcanoclastic, (e) phenocryst of amphibole in altered context of lava flow, (f) amphibole phenocryst along with clinopyroxene grains in lava flow.

Analytical method

The XRF, ICP-MS and ICP-AES analyses were performed at the Centre de Géochimie de là Surface, Strasbourg, France. Approximately 1 gram of agate-crushed sample was dried first at 110 °C and then held at 1000 °C for three hours. 100 mg of the charred sample was mixed with 750 mg of lithium tetraborate in a graphite furnace. This mixture was melted at 1000 °C for 30 minutes. After cooling, the obtained bead was directly dissolved in a furnace by a solvent containing diluted nitric acid and glycerin. The final solution was 4 g/L or 100 mg of rock in 25 ml of solvent. After filtering, the solution was directly entered into the ICP-AES. The same solution, diluted 10 times, was then introduced into the ICP-MS. The precision of ICP-AES is ±10 % while ICP-MS yields a precision of ±5 %. Major elements along with Ni, Cr, V, Sc, Y, Zr, Ba, Sr were measured by ICP-AES, whereas other trace and rare earth elements were determined using ICP-MS. Table 1 includes the results of major, trace and rare earth element analyses of the basaltic rocks of the Chaldoran region.

Geochemistry

Based on field studies and available geochemical data, the studied rocks of the Chaldoran region can be divided into three groups. The first group is ophiolite (pillow lava, gabbro and dykes that are rooted to ultramafic rocks). The second group includes ophiolitic-related basaltic fragments and breccias. The third group is related to Eocene basaltic lavas.

Ophiolitic rocks (first group) show a wide range of major element content. This group of rocks contains <56 wt% SiO_2 (44.5-55.7 wt %) and is characterized by low to high amounts of Al2O3 (5.4-18.3 wt %) and MgO (2.7-21 wt %) along with low to moderate amounts of TiO2 (0.06-1.1wt %). Generally, the content of TiO2 in a mafic series attributed to a subduction zone is less than 1 wt%. In the logarithmic diagram of Nb/Y versus Zr/TiO2 all of these rocks are plotted in a sub-alkaline basalt field (Fig. 4)[15]. These series of basalts show different REE patterns and are enriched with LREE (Fig. 5a)[16]. These rocks show enrichment with K, Sr, U and a depletion of Nb, Ta (Fig. 5b)[16]. These geochemical characters are related to the rocks of the SSZ. In the Ta/Th/Hf diagram, these samples are plotted in the calc- alkaline volcanic arc basalt series (Fig.6)[17].

The SiO2, Al2O3, MgO and TiO2 in the second group of rocks varies from 54 to 61, 13.4 to 17.3, 1.3 to 3.7 and 0.4 to 1 wt% respectively. On the basis of logarithmic diagrams (Nb/Y versus Zi/TiO2), ophiolitc-related mafic fragments are plotted in the sub-alkaline basalts and andesite field (Fig. 4)[15]. The pattern of REE in these samples is highly different and marked by enrichment of light rare earth elements (Fig. 5c)[16]. All samples displayed marked negative HFSE (Nb, Ta) and positive LILE anomalies typical of subduction-related magmas (Fig. 5d). Rocks in the Ta-Th-Hf diagram are plotted in the calc alkaline volcanic arc basalt field (Fig. 6)[17].

In the third group, the SiO2 content of samples varies from 33 to 61 wt%, the Al2O3 content varies from 10.5 -18.6 wt% and the MgO content varies from 1.4-6.9 wt%. The TiO2 content of these samples (1.1 to 1.8) is high in comparison to the calc-alkaline series. The high TiO2 content of these rocks is in agreement with the existence of TiO2 bearing minerals such as titanite and titanomagnetite. In the Nb/Y versus Zi/TiO2 diagram these rocks are plotted in the alkaline basalt field (Fig. 4)[15]. In the chondrite normalized diagram, these rocks display enrichment in LREE (Fig. 5e)[16]. In comparison to the calc-alkaline series, these types of basalts are characterized by positive anomalies of HFSE and LILE (Fig. 5f)[16]. The pattern of REEs and the multi element diagram of these rocks is similar to OIB-like basalt. On a Th-Hf-Nb diagram, these rocks plot in the plate alkaline basalt field (Fig. 6)[17].

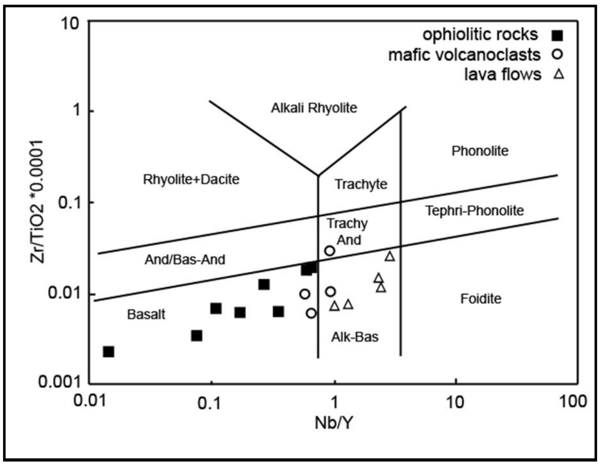

Fig. 4. Zr/TiO2 versus Nb/Y diagram classification diagram (Pearce, 1996) [14] for the studied rocks.

Discussion and conclusions

The chemical compositions of basaltic rocks is very important in distinguishing the tectonic setting, petrogenesis and magmatic evolution of rocks. Differences in trace element content of various rocks may be associated with differences in sources. The overall enrichment of LILE and LREE with negative Nb, Ta anomalies is a feature of subduction-related magmas and is commonly attributed to a mantle source which has been previously enriched in LILEs over HFSEs by metasomatic activity of fluids derived from the subduction slab or sediments (e.g. [18, 19]).

As shown in figure 7[16, 18, 20, 21], some elements, for example, Th ,Nb and Yb, can be used to identify sources with respect to processes such as crustal enrichment, crustal assimilation and contamination. The Yb is much less incompatible than Th and Nb which are strongly incompatible and show enrichment and depletion in subduction zones, respectively. In figure 8, samples associated with ophiolite show a calc-alkaline affinity and plot in the continental volcanic arc field. These rocks are influenced by fluids in a continental subduction environment. However, the samples that are not related to ophiolites (Eocene basalts) are plotted in the MORB-OIB field.

The HREE distribution patterns in chondrite-normalized diagrams for the studied samples are relatively flat (Fig. 5)[16]. This indicates that HREE fractionations are weak, and do not support the presence of garnet in the mantle source region (Aydin et al., 2008)[19]. Instead, the patterns are consistent with the presence of spinel in the mantle source. The average Ce/Pb ratios of the studied ophiolitic basalts (3.7-19.3) and volcanicalstics (5.7-16.7) are considerably different from those of oceanic basalts (~25; Hofmann, 1988)[22]. It more likely suggests that these rocks are not derived from normal asthenospheric mantle. Moreover, Bradshow and Smith (1994)[23] and SMITH et al. (1999)[24] mentioned that, since HFSE are depleted in the lithospheric mantle relative to the LREE, high Nb/La ratios (~>1) indicate an OIB-like asthenospheric mantle source for basaltic magmas, and lower ratios (~ <0.5) indicate a lithospheric mantle source. The Nb/La and La/Yb ratios of the most studied rocks suggest a mixed lithospheric-asthenospheric mantle source in figure 8[24, 25, 26].

The trace element pattern of Chaldoran's Eocene basalts in the case of HFSE and LILE abundance are different from Eocene rocks that were studied by Dilek et al. (2010)[3] in eastern Turkey, Azarbayjan and various regions of Iran. The studied rocks by Dilek et al. (2010)[3] show depletion in HFSE and enrichment in LILE and the source of the mentioned rocks is attributed to metasomatized mantle related to a subduction zone environment. However, the Chaldoran Eocene rocks not only show no depletion in HFSE, they are also marked by a lesser enrichment in LILE in comparison to Dilek et al. (2010)[3] samples. Based on this information, it is clear that the genesis of Chaldoran Eocene rocks is incompatible with the model proposed by Dilek et al. (2010)[3] for the Cenozoic rocks in this area.

Reaganand Gill (1989)[27] showed that incompatible trace element enrichment in magma can be inherited from small-degree partial melting of an OIB-like mantle source.

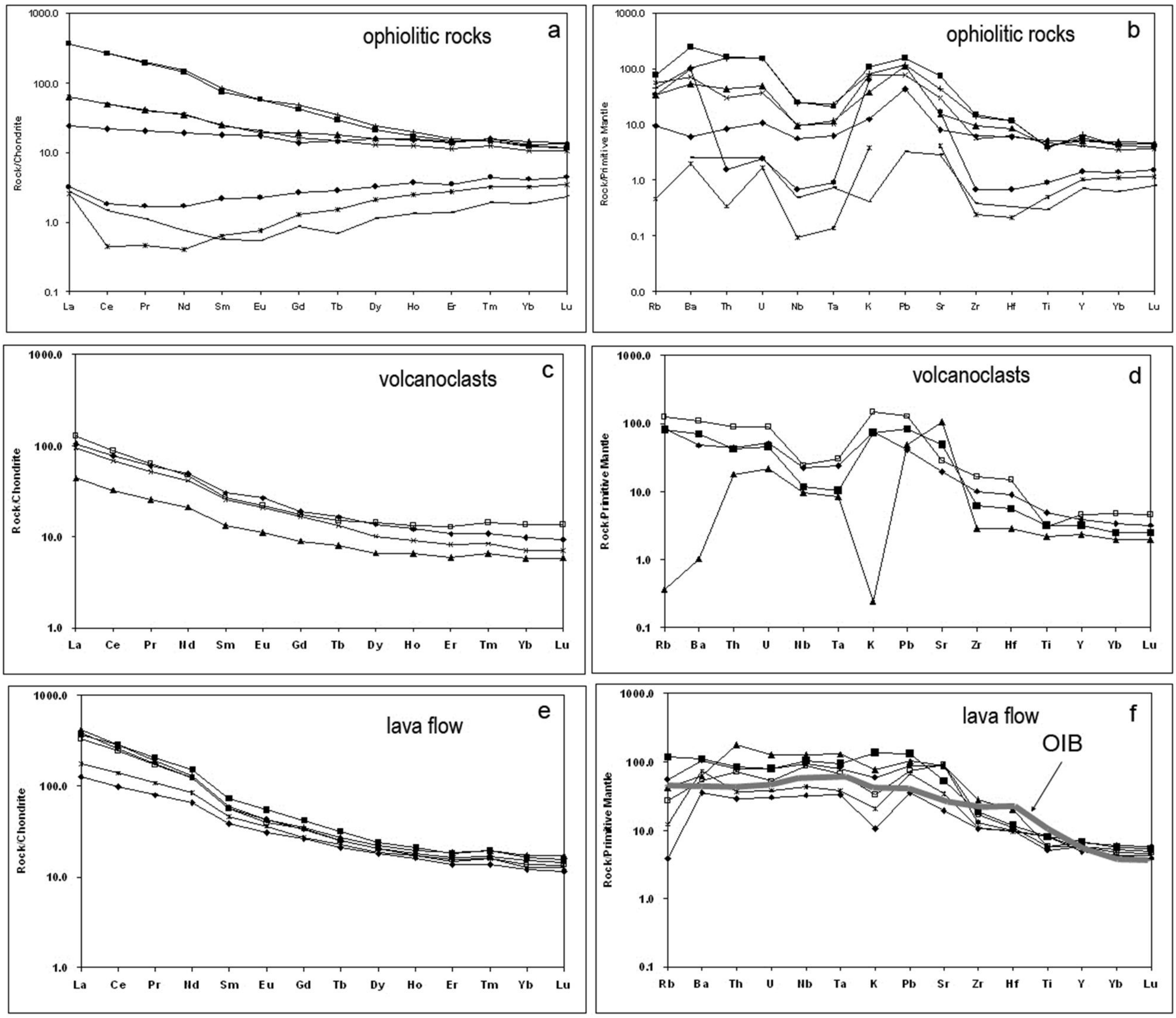

Fig. 5. Chondrite-normalized REE (right) and primary Mantle–normalized (left) diagrams for the basaltic rocks from Chaledran region (normalizing values are from Sun & McDonough 1989)[15].

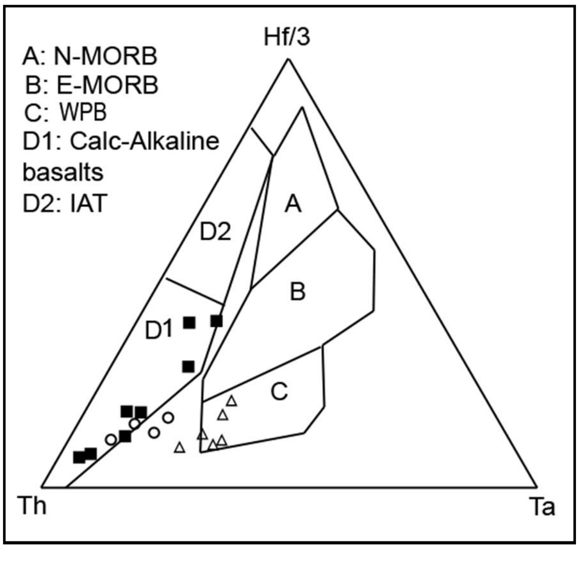

Fig. 6. Tectonomagmatic discrimination Hf-Nb-Th diagram for the studied rocks from Wood (1980) [16]. Symbols are as in Fig. 4.

Table 1: Whole rock chemical analyses of the basaltic rocks from Chaldoran region.

Rock type	pillow	pillow	metabasalt	microgabbro	gabbroic dyke	pegmatite gabbro	diabasic dyke	diabasic dyke
Sample No.	M06-113	M06-140B	M06-6	M06-55	M06-180	M06-21	M06-177	M06-170
SiO2	46.5	49.6	55.7	53.1	48.8	49.7	48.7	44.5
Al2O3	17.5	18.3	12.2	14.5	3.76	17.9	12.6	5.39
MgO	7.64	5.07	2.69	3.94	18.2	9.73	6.90	21.1
CaO	11.1	8.65	7.92	6.96	16.0	8.01	11.7	15.7
Fe2O3	9.79	9.04	7.53	7.36	9.82	5.43	10.28	6.81
MnO	0.135	0.146	0.100	0.123	0.164	0.109	0.142	0.114
TiO2	1.11	0.859	0.885	1.03	0.109	0.194	0.820	0.064
Na2O	2.49	2.62	3.78	2.61	0.096	2.16	3.64	0.084
K2O	0.379	3.30	1.15	2.30	0.118	1.91	2.46	0.012
P2O5	0.138	0.714	0.237	0.266	-	0.006	0.588	-
Sum	99.23	100.31	98.92	98.90	98.60	99.11	99.99	98.55
Sr	171	1570	330	624	87	355	921	61
Ba	42.2	1719	374	511	14	690	741	18
V	207	252	129	232	436	239	262	148
Ni	107	28.2	12	16.5	525	248	66.9	555
Cr	257	55.8	22	0.71	1294	164	210	1711
Zn	88.4	90.9	126	66.3	66	18	79.8	42
Cu	82.3	216	20	85.1	201	68	84.3	23
Sc	33.7	17.8	18	23.3	81	30	33.6	37
Y	23.3	26.7	24	19.0	5	7	29.7	3
Zr	68.8	169	106	63.4	3	7	158	4
Co	44.1	28.0	9.07	18.8	55.2	30.9	37.3	45.6
Rb	6.15	49.6	21.6	36.4	0.293	21.9	28.9	-
Nb	4.09	17.9	6.78	7.07	0.067	0.483	17.7	0.345
Sn	2.44	2.11	0.761	1.346	-	-	0.931	-
Sb	0.083	0.106	0.368	0.114	0.038	-	0.148	0.026
Cs	0.140	0.318	-	0.767	-	1.42	0.222	-
La	5.78	86.4	14.9	15.2	0.611	0.749	87.7	0.633
Ce	13.3	163	30.1	30.0	0.272	1.11	164	0.890
Pr	1.95	18.4	3.87	3.80	0.043	0.160	18.9	0.104
Nd	9.04	66.7	16.2	16.4	0.192	0.799	70.7	0.345
Sm	2.72	11.4	3.80	3.67	0.098	0.331	12.8	0.086
Eu	1.008	3.36	1.126	1.16	0.044	0.129	3.35	0.031
Gd	2.88	8.54	3.921	3.35	0.263	0.545	9.83	0.177
Tb	0.541	1.09	0.667	0.543	0.056	0.107	1.30	0.026
Dy	3.94	5.36	4.06	3.25	0.528	0.827	6.20	0.291
Ho	0.860	0.965	0.893	0.705	0.140	0.209	1.12	0.077
Er	2.27	2.39	2.37	1.88	0.461	0.572	2.58	0.229
Tm	0.407	0.382	0.402	0.312	0.081	0.112	0.370	0.048
Yb	2.22	2.09	2.42	1.76	0.539	0.692	2.04	0.313
Lu	0.336	0.294	0.346	0.267	0.087	0.113	0.292	0.060
Hf	1.83	3.61	2.66	1.89	0.067	0.207	3.68	0.104
Ta	0.261	0.857	0.482	0.413	0.006	0.037	0.941	0.030
W	0.88	0.68	0.914	0.708	0.374	0.113	0.500	0.422
Pb	3.13	11.3	7.98	5.62	-	-	8.45	0.233
Th	0.702	13.9	3.77	2.53	0.029	0.134	13.1	0.211
U	0.228	3.21	1.03	0.763	0.035	0.051	3.23	0.055

Table1: (continued)

Rocktype	andesite	andesite	andesite	andesite	lava flow	Lava flow	Lava flow	lava flow	lava flow	basaltic dyke
sample_No.	M06-161	M06-162	M06-168	M06-70	M06-137	M06-151	M06-112	M06-134A	M06-134B	M06-144
SiO2	54.5	61.2	55.1	54.0	42.5	40.2	48.0	42.9	43.1	44.9
Al2O3	17.3	15.2	13.4	15.6	14.7	17.2	18.6	16.5	15.0	15.6
MgO	2.49	1.94	1.37	3.73	6.89	5.71	2.77	4.79	6.86	5.86
CaO	8.74	3.71	16.6	8.67	12.0	11.2	9.59	9.89	13.7	10.7
Fe2O3	5.91	5.73	7.73	8.64	11.3	11.0	9.31	12.7	12.0	9.47
MnO	0.081	0.093	0.092	0.128	0.180	0.233	0.206	0.288	0.210	0.216
TiO2	1.06	0.668	0.477	0.691	1.78	1.28	1.23	1.74	1.76	1.11
Na2O	4.62	3.61	0.000	2.35	3.82	5.26	3.34	2.13	2.62	4.65
K2O	2.28	4.45	0.007	2.23	0.320	0.998	2.30	4.11	0.629	1.79
P2O5	0.483	0.251	0.237	0.346	0.335	0.626	0.537	1.12	0.533	0.570
Somme	100.01	98.74	99.08	99.33	98.71	99.01	99.66	99.66	99.90	100.01
Sr	417	593	2209	1034	409	1923	1873	1096	711	1849
Ba	338	774	7.15	489	247	378	441	764	504	727
V	163	96.6	178	249	244	234	127	162	257	151
Ni	16.9	2.47	22.1	13	89	31	9.71	9.94	84.7	107
Cr	0.401	-	10.5	9	128	12	-	2.63	119	230
Zn	32.6	49.4	21.2	72	102	109	96.6	120	106	88.4
Cu	62.9	27.1	4.80	129	107	83	12.3	33.2	86.5	42.3
Sc	12.5	9.33	14.6	26	30	8	6.19	4.39	24.2	15.1
Y	17.5	21.1	10.7	15	22	26	29.9	31.0	24.0	25.9
Zr	113	186	31.7	69	120	186	313	213	121	144
Co	15.6	13.0	9.35	19.3	42.5	27.9	16.7	31.3	47.6	29.7
Rb	53.0	78.8	0.233	51.6	2.41	17.0	27.0	74.5	7.79	34.6
Nb	15.9	17.5	6.81	8.29	23.4	61.0	89.4	74.8	31.1	66.0
Sn	1.42	1.76	0.544	0.379	0.474	0.299	1.76	2.60	2.83	0.796
Sb	0.233	0.223	0.284	0.133	0.185	0.047	0.241	0.277	0.173	0.088
Cs	0.259	0.316	0.056	0.401	0.063	-	0.095	0.883	0.362	0.316
La	25.1	30.4	10.5	22.5	30.3	78.0	98.4	87.2	41.4	91.9
Ce	47.9	53.8	20.0	41.3	60.0	148	176	173	84.3	159
Pr	5.73	5.99	2.46	4.91	7.55	16.5	18.1	19.7	10.3	17.1
Nd	22.9	22.0	9.96	19.4	30.5	58.3	60.3	71.3	39.5	58.5
Sm	4.62	4.18	2.02	3.90	5.94	8.74	9.03	11.1	7.17	8.67
Eu	1.55	1.26	0.646	1.199	1.795	2.299	2.51	3.17	2.10	2.48
Gd	3.94	3.57	1.84	3.368	5.441	6.865	7.13	8.60	5.55	6.82
Tb	0.614	0.563	0.301	0.494	0.788	0.935	1.02	1.19	0.849	0.935
Dy	3.50	3.65	1.70	2.57	4.59	5.15	5.62	6.06	4.70	5.20
Ho	0.691	0.756	0.372	0.515	0.920	0.969	1.11	1.21	0.962	1.04
Er	1.79	2.12	0.984	1.36	2.26	2.56	3.06	3.03	2.45	2.72
Tm	0.277	0.363	0.166	0.211	0.350	0.408	0.498	0.500	0.406	0.426
Yb	1.67	2.31	0.981	1.22	2.05	2.33	2.97	2.78	2.14	2.56
Lu	0.235	0.343	0.148	0.181	0.289	0.343	0.431	0.398	0.317	0.362
Hf	2.78	4.53	0.883	1.70	3.04	3.39	6.30	3.64	2.93	3.06
Ta	0.983	1.24	0.339	0.430	1.35	2.77	5.34	3.85	1.59	3.32
W	0.786	1.01	0.705	0.577	0.199	1.46	0.664	1.07	0.85	0.497
Pb	2.87	9.02	3.46	5.93	2.45	5.34	7.24	9.32	4.82	6.21
Th	3.77	7.57	1.53	3.65	2.46	6.10	15.1	7.13	3.10	6.59
U	1.07	1.89	0.454	0.955	0.637	1.09	2.64	1.69	0.811	1.63
Ti	6380.68	4005.74	2859.78	4144.03	10652.3	7648.13	7388.33	10411.5	10531.6	6645.87
K	18892.6	36930.3	60.9548	18532.4	2657.73	8281.82	19052.0	34091.3	5221.43	14857.6
P	2108.29	1095.16	1033.45	1508.86	1459.98	2732.25	2345.24	4904.47	2326.43	2487.19

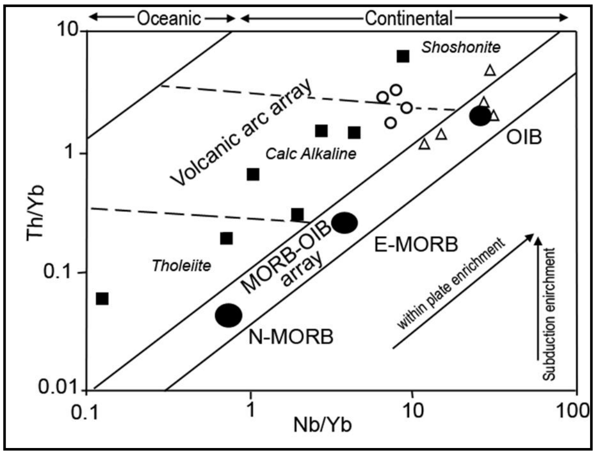

Fig. 7. Th/Yb versus Nb/Yb diagram from Pearce (1983)[17], Pearce and Peate (1995)[19], and Stern et al. (1995)[20]. N-MORB, E-MORB, and OIB from Sun and McDonough (1989)[15]. Symbols are as in Fig. 4.

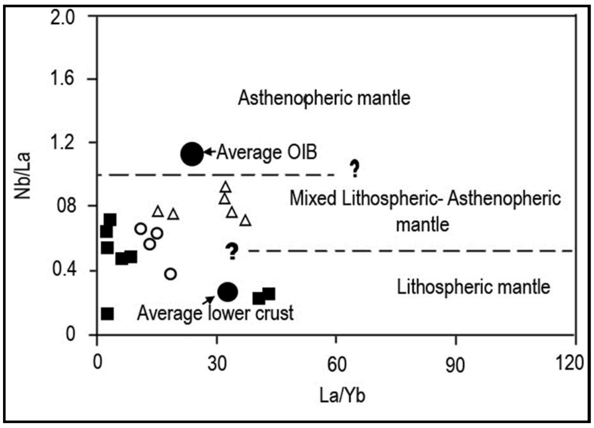

Fig. 8. Nb/La versus La/Yb variation diagram. Average OIB is after Fitton et al. (1991)[24] and average lower crust is after Chen and Arculus (1995)[25]. Dashed lines separating fields of the asthenospheric, lithospheric and mixed mantle are plotted based on data given in Smith et al. (1999)[23]. Symbols are as in Fig. 4.

Furthermore, numerical modeling by McKenzie (1989)[28] and Tainton and McKenzie (1994)[29] demonstrated that melts emanating from mantle upwelling can be enriched with volatiles and incompatible elements. It seems more likely that partial melting of an OIB-like mantle source generated magma with trace element contents similar to those of the Chaldoran basalts. Overall LREE-enriched and moderate to high La/Yb (14.8 to 35.8) rations of most Eocene Chaldoran rocks indicate that these rocks were derived as a result of lower degrees of partial melting than the ophiolitic rocks that have low La/Yb rations. Therefore, the geochemical constraints of Eocene Chaldoran volcanic rocks are in line with a local hotspot/plume origin.

The basement of the study area shows remnants of Cretaceous oceanic crust. According to Khalatbari-Jaffari et al. (2003, 2004)[6,7], there was an oceanic basin in the southern part of the Khoy study area during the Mesozoic era (Upper Triassic to Cretaceous). They believe that subduction of this oceanic basin beneath the continental margin in the northern basin (Azarbaijan block) started during the Jurassic period and accumulated subducted slabs in the northern continental margin, finally forming a metamorphic complex of east Khoy ophiolite. The study area is comparable to the active Iranian continental margin (Azarbaijan block) from the Upper Triassic to early Cretaceous. Due to a changing Neo-Tethys subduction regime in the Middle Cretaceous (Agard et al., 2006)[30], a wide oceanic basin formed in the western portion of the primary Khoy basin during the middle Cretaceous which showed an extension trend different from the primary basin in the Khoy area. This rifting continued to the active continental margin and formed an oceanic basin in the Chaldoran region during the Middle Cretaceous. This oceanic basin was situated between the Azarbaijan block and portions of the Sanandaj-Sirjan block. The chemical characters of ophiolitic rocks confirm that the Chaldoran oceanic crust formed in a subduction continental setting and that the basin was part of the east Turkish and Khoy Upper Cretaceous basins. The absence of a transition zone, chromitite lenses and sheeted dykes, along with a thin crustal sequence, probably suggests that the spreading rate in this seaway was slow during its formation during the Cretaceous period.

Due to the wide expansion of ophiolitic rocks, lack of Cenozoic magmatic evidence of a SSZ nature and spreading of a large volume of Eocene-Paleocene sediments in the Chaldoran region, presenting a unique view of the closing or narrowing of this basin is difficult. Based on our study it is highly likely that magmatism occurring after the Cretaceous period isn't related to the spreading or closing of the Chaldoran oceanic area and magmatism after the Cretaceous is not related to ophiolites. Due to the presence of volcaniclasts and cold breccia with different compositions (acidic to basaltic) and their SSZ natures, the narrowing of this basin occurred along with inter-oceanic subduction. However, questions remain considering that they show a continental arc nature. The Chaldoran region is situated in an active continental margin of the Iranian block (Azarbajan block) with the Neo-Tethys subducting underneath it. A changing subduction regime of the Neo-Tethys happened during the Middle Cretaceous (Agard et al. 2006)[30] along with activation NW of the active continental subduction regime of Iran. Hence, it is possible that volcaniclastic rocks with a continental arc nature may be the result of magma ascending into oceanic islands. Magmas from these rocks were generated as a result of the partial melting of the mantle wedge onto the Neo-Tethys subducted slab and sub-oceanic crust of the active Chaldoran Ocean during the Upper Cretaceous.

Cenozoic magmatic activity started with the generation of alkaline basaltic lavas in the upper Eocene that are the result of uprising mantle plumes. They do not show any SSZ rock characteristics. The hault of spreading in the Chaldoran basin and some other closed basins in the inner part of Iran such as Nain-Baft during the Upper Cretaceous (Ghazi et al. 2011a, b)[31, 32] may have led to changes in the Neo-Tethys subduction régime in northwestern Iran. These events led to the breaking off of the subducted Neo-Tethys crust in the sub-oceanic crust of the Chaldoran hinterland (as a part of an active continental margin of Neo-Tethys) and the development of an asthenospheric window.

The mixing of ascending asthenospheric magma and with lithosphric magma led to the formation of Eocene alkaline magmatism in the Chaldoran hinterland. This magma in the course of crossing through ophiolitic rocks filled cracks along the faults in the ophiolitic assemblage. Therfore, some of these alkaline magmas penetrated and cooled among ophiolitic formations which is why according to some studies such as RAHGOSHAY et al. (2008) and Pormohssen et al. (2010)[6], ophiolitic rocks are divided into two groups. They believe that these alkaline and OIB-like rocks formed by chance since these rocks weren't generated during spreading oceanic crust and are not genetically related to ophiolitic rocks. During the Oligocene, the sedimentation conditions were weaker than during the Miocene and the Cenozoic sediment that covered major parts of the Cretaceous oceanic crust is mainly Miocene in age.Furthermore, numerical modeling by McKenzie (1989)[28] and Tainton and McKenzie (1994)[29] demonstrated that melts emanating from mantle upwelling can be enriched with volatiles and incompatible elements. It seems more likely that partial melting of an OIB-like mantle source generated magma with trace element contents similar to those of the Chaldoran basalts. Overall LREE-enriched and moderate to high La/Yb (14.8 to 35.8) rations of most Eocene Chaldoran rocks indicate that these rocks were derived as a result of lower degrees of partial melting than the ophiolitic rocks that have low La/Yb rations. Therefore, the geochemical constraints of Eocene Chaldoran volcanic rocks are in line with a local hotspot/plume origin.

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