Origin and Formation Process of Gabbro and Diorite Xenoliths in the Northern Part of Uromieh-Dokhtar Magmatic Belt, NW Iran: Constraints from Mineral and Whole-Rock Chemistries

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ABSTRACT

Gabbro and diorite xenoliths occur within the alkaline trachyandesites in the northern part of Uromieh-Dokhtar magmatic belt (NW Iran). The host volcanic rocks are located on the upper Miocene sediments; therefore, they are probably Plio-Quaternary in age. The texture of xenoliths is granular, and the gabbro xenoliths are classified into two groups based on their petrographic characteristics and mineral chemistry. Based on chemical analyses, xenoliths show a subalkaline to slightly alkaline magmatic trend. Thermobarometric results indicate that gabbro and diorite xenoliths formed within the middle crustal level. All xenolith samples are light rare earth elements (LREEs) enriched and depleted in Ta, Zr, and Nb. These features can be explained by the subduction of the Neotethys. The LREE enrichment as well as negative Th, Ti, Nb, and Zr anomalies and positive Pb anomaly are in agreement with an origin through partial melting of an enriched mantle previously affected by metasomatism processes linked to mantle wedge-derived melts (fluids). Different REE concentration levels and patterns for Group 1 gabbro, Group 2 gabbro, and diorite xenoliths point out that their parental melts were not the same. Those forming the Group 1 gabbro xenoliths originated from melting of garnet mantle lherzolite, whereas the parental melts from Group 2 gabbro and diorite xenoliths originated from melting of spinel-garnet mantle lherzolite. The $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios and chemical composition of minerals are different for host volcanic rocks and xenoliths; therefore, we may conclude that the parental melts of the different groups of xenoliths and host volcanic rocks differ in composition.

Online enhancements: color figures, supplemental material.

Introduction

Many researchers attribute the igneous rocks of Uromieh-Dokhtar magmatic belt to Neotethys subduction (e.g., Hassanzadeh 1993; Mohajjel et al. 2003). The presence of potassium-rich lavas in this area is important, so these rocks occur in a variety of tectonic settings including continental convergent margins (Gill et al. 2004), and syn- to postcollisional tectonic settings as in Anatolia and Tibet (Conticelli et al. 2009, 2013; Luo et al. 2009; Prelevic et al. 2010; Deng et al. 2012). The presumed Plio-Quaternary volcanic rocks represent the latest magmatic event in this

region. These volcanic rocks contain various xenoliths, including pyroxenite, hornblendite, amphibolite, gabbro, and diorite. The most important studies in this area were by Ahmadzadeh (2010), focusing on the Galleban area; by Khezerlou (2006) and Parkani (1998), focusing on the northwest of Marand and Zanjireh areas, respectively; and by Khezerlou et al. (2017), focusing on pyroxenite xenoliths and their host volcanic rocks. The previous geochemical and tectonic studies of volcanic rocks have demonstrated their dominant alkaline affinity and their formation in a postcollisional environment (e.g., Parkani 1998; Khezerlou 2006; Ahmadzadeh 2010).

In their study of pyroxenite xenoliths from the Uromieh-Dokhtar magmatic belt, Khezerlou et al.

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(2017) proposed that their magmas originated from enriched mantle. Gabbros and diorites are the most abundant xenolith types in the volcanic rocks from the studied area; however, the research focusing on xenoliths have been scarce. Alkaline magmas move rapidly to the surface after production, and they can carry xenoliths, including gabbroic and dioritic xenoliths, to the surface (Pearson et al. 2003). Previous studies have reported several examples of gabbro xenoliths formed from alkaline magmas in different locations worldwide (e.g., Preston and Bell 1997; Pedersen and Sørensen 2003; Al-Mishwat and Nasir 2004; Egorova et al. 2006; Tollan et al. 2012). The study of the geochemical and isotopic characteristics of gabbro and diorite xenoliths may help to better understand and constrain the nature and evolution of magmatic activities from the northern Uromieh-Dokhtar magmatic belt.

In this article, bulk rocks and mineral chemistry analysis as well as Nd and Sr isotopic data for gabbro and diorite xenoliths from the northern Uromieh-Dokhtar magmatic belt are presented. Using these data, we evaluate the origin and history of these xenoliths. In addition, the chemical compositions of minerals of xenoliths and host volcanic rocks are investigated, to better understand the link between

the magmatic sources of gabbro and diorite xenoliths and host volcanic rocks, and finally, the thermobarometry of xenoliths is also carried out.

Geological Setting

The study area is located in the northern Uromieh-Dokhtar magmatic belt (NW Iran; fig. 1), part of the Tertiary-Quaternary magmatic belt of Iran (Stöcklin 1977). Neotethys subduction beneath central Iran during the Upper Cretaceous and Paleogene and the collision of the Iran and Arabia Platforms created the following four structural zones in Iran: the High Zagros belt, the Sanandaj-Sirjan zone, the Uromieh-Dokhtar magmatic belt, and the Zagros folded zone (Alavi 2004; fig. 1). The study area is located in the northernmost part of the Uromieh-Dokhtar magmatic belt, characterized by magmatic activity that started during the Late Cretaceous and active from Eocene to Quaternary. The peak of magmatism in this zone occurred during the Eocene (Farhoudi 1978; Emami 1981; Alavi 2004). It temporarily paused and experienced a lack of volcanic activity during the middle Oligocene to middle Miocene. The volcanic activity started again in the upper Miocene to Plio-Quaternary (Omrani 2008).

Figure 1. The four main tectonic units of the Zagros Orogenic Belt (after Alavi 1994; modified from Ghasemi and Talbot 2006). A color version of this figure is available online.

Eocene volcanism consists mainly of calc-alkaline compositions, and those rocks lack feldspathoid minerals. Upper Miocene to Plio-Quaternary magma has alkaline and shoshonitic tendency, and in this case, rocks contain feldspathoid minerals (Jung et al. 1976; Berberian et al. 1982).

Alkaline magmas move rapidly to the surface after production, and they can carry xenoliths, including gabbroic and dioritic xenoliths, to the surface (Pearson et al. 2003). Alkaline volcanic rocks have a limited extent and are commonly located near the main faults. The tectonic maps show that the volcanic rocks from the study area are located in between the main faults of the region including the North Tabriz, Misho, and Tasuj faults to the south and the Gharagoz and South Divan Daghi faults to the north (fig. 2). Alkaline volcanic rocks have also been reported from the Eslamieh Peninsula (Mora-

dian et al. 1997; Moayyed et al. 2008; Shafaii Moghadam et al. 2013), and Shafaii Moghadam et al. (2013) estimated their age at 11 My (late Miocene).

The volcanic rocks from the northwest of Marand lie on upper Miocene sediments (fig. 3), so they are probably related to Plio-Quaternary magmatic activity in the region. Field surveys showed that a wide range of volcanic rocks outcrop in this area (fig. 4). These rocks are leucite basanite, leucite tephrite, tephrite, trachyandesite, andesite, trachydacite, and dacite. Xenoliths occurring in some volcanics from this area are gabbros, diorites, hornblendites, lamprophyres, hornfelses, amphibolites, and pyroxenites. Hornblendite, gabbro, and pyroxenite xenoliths are the most abundant. Xenoliths are hosted by trachyandesites and basalt trachyandesites, but gabbro and diorite xenoliths are found only within the trachyandesites. Lamprophyric dikes also occur in

Figure 2. Tectonic map of the investigated area showing the location of the outcrop of xenolith-bearing volcanic rocks. Gh.G.F. = Gharagoz fault; N.M.F. = North Misho fault; N.T.F. = North Tabriz fault; S.D.F. = South Divan Daghi fault; S.M.F. = South Misho fault; Ts.F. = Tasuj fault.

Figure 3. Outcrop of host volcanic rocks and upper Miocene sediments from the study area. A color version of this figure is available online.

the study area. Among them, we can mention the lamprophyre from Eslamieh Peninsula (Moayyed et al. 2008) and from Gikhlar of Marand (Moazzen et al. 2003). Some lamprophyric dikes between 9 and 11 Ma (late Miocene) in age also occur in the study area in Sorkheh (Aghazadeh et al. 2015). In the south of the study area, in Misho Mountain, mafic and ultramafic rocks occur, including gabbro, anorthosite, olivine gabbro, peridotite, and pyroxenite. Sacani et al. (2013) reported petrological, geochemical, and geochronological data from the Misho Mafic Complex. The U-Pb zircon age of a leucogabbro dike has revealed that the igneous emplacement age of these rocks is 356.7 ± 3.4 Ma. As a result, they ascribe the Misho Mafic Complex to the Paleotethys. In the Misho area, there are also acidic rocks including monzogranite, granodiorite, diorite, and granite. These rocks (acidic rocks) intersect the Misho Mafic Complex (Shahzaidi 2013).

Petrography

Host Volcanic Rocks. The volcanic series from the study area consists of SiO_2 -undersaturated rocks (e.g., leucitite, leucite basanite, leucite tephrite, basanite, and tephrite) to SiO_2 -saturated rocks (e.g., trachyandesite, andesite, trachydacite, and dacite). The gabbro and diorite xenoliths occur within the trachyandesite rocks. The trachyandesites show textures that are porphyritic (fig. 5a) and microlithic porphyritic (fig. 5b).

Andesine plagioclase is the most abundant mineral phase in these volcanic rocks, ranging from 45% to 50% of total modal content. Their size varies from very fine-grained microliths (less than 0.1 mm in size) occurring within the matrix of phenocrysts up to crystals of 1.2 mm in size. In some cases, the variation in size is transitional, resulting in the seriate texture (fig. 5b). Most of the plagioclases display Carlsbad and polysynthetic twins, and some of them have a sieve texture (fig. 5a, 5b) or even zoning (fig. 5a). Clinopyroxene (diopside) constitutes 8%–10% of the samples. This mineral occurs as phenocryst but not as microlith in the matrix. Diopside phenocrysts are euhedral to semieuhedral and range in size from 0.2 to 1.5 mm. The minor mineral phases of these rocks include amphibole, biotite, K-feldspar, apatite, and opaque minerals associated with glass in some samples. Euhedral and semieuhedral brown hornblende ranging in size from 0.2 to 1 mm also occurs. This mineral makes up to 4%–5% of the modal content of some samples, and some crystals display opacitization (fig. 5a). Another minor mineral is biotite, which could make up to 4%–6% of the mode. Its size ranges from 0.3 to 0.8 mm, and some crystals also display opacitization (fig. 5a). Apatite and opaque minerals smaller than 0.5 mm (1%–2% of the mode) commonly occur.

Gabbro Xenoliths. Hand specimens of these xenoliths are light gray in color. Bright plagioclase and dark green clinopyroxene can be clearly identified in such samples. These rocks show granular textures,

Figure 4. Geological map of the investigated area (simplified from the 1 : 100,000 geological maps of Marand, Jolfa, Gharezyaadin, and Tasuj; modified after Khezerlou et al. 2017).

and their main constituent minerals include plagioclase, clinopyroxene, and amphibole. Based on the petrographic characteristics and mineral chemistry, these xenoliths are divided into two groups. Sub-euhedral plagioclases varying in size from 0.5 to 2 mm constitute 40%–45% of the modal content of rocks from these two groups, and most of them display Carlsbad and polysynthetic twins (fig. 5b, 5c). The plagioclase of Group 1 is bytownite, whereas that of Group 2 is labradorite. Subeuhedral clinopyroxenes ranging in size from 0.3 to 1.5 mm make up about 15%–20% of the modal content of those two groups of gabbro xenoliths. Clinopyroxenes of Group 1 are slightly darker than those of Group 2. Tiny opaques are observed within some of the clinopyroxenes (fig. 5c). Amphiboles in the two groups of gabbro xenoliths

are euhedral and semieuhedral and have brown polychroism. They range in size from 0.3 to 1.5 mm and make up about 10%–20% of the modal contents. Within some of the amphiboles, tiny crystals of clinopyroxene, plagioclase, opaque, and apatite occur (poikilitic texture; fig. 5c).

Biotite could be considered a main constituent mineral (10%–12% of the modal content) in some of these xenoliths, but in other cases it is a minor phase (less than 8%). Most biotites are subeuhedral and range from 0.2 to 1.2 mm in size. Other minor minerals include apatite and opaque. Apatites mostly occur as euhedral crystals of less than 0.2 mm in size and make up about 0.5%–1% of the mode of samples. The apatite modal content in Group 1 is higher than in Group 2. This mineral occurs as inclusions within

Figure 5. Photomicrographs of host volcanic rocks and xenoliths from the study area: *a*, Plagioclase (Plg), clinopyroxene (Cpx), amphibole (Amph), and biotite (Bt) phenocrysts in trachyandesite, crossed polars (XPL). *b*, Plg phenocrysts in trachyandesite and gabbro with Amph, Cpx, and Plg, XPL. *c*, Gabbro with Amph, Cpx, Plg, and apatite (Ap), XPL. *d*, Metasomatized gabbro with Amph, Bt, Plg, Cpx, and opaques, XPL. *e*, Plg, Amph, Bt, Cpx, and opaques in metasomatized gabbro, XPL. *f*, *g*, Diorite with Amph, Cpx, mica, and Plg, XPL. *h*, Metasomatized diorite with Amph, Bt, Cpx, Plg, and opaques, XPL. A color version of this figure is available online.

amphibole but also as interstitial crystals. Opaque minerals occurring in all samples are variable in size from one sample to another and commonly make up less than 2% of the mode of rocks. Some of the gabbro xenoliths display clinopyroxenes rimmed by amphiboles (fig. 5*d*, 5*e*). In those samples, some of the plagioclases are decomposed into clay minerals.

Diorite Xenoliths. These xenoliths are also bright gray in hand specimen samples and are less abundant than gabbroic xenoliths.

The texture of this group is granular, like that of gabbro xenoliths. Plagioclase and amphibole are the main constituent minerals. Plagioclases are subeuhedral, and some show Carlsbad and polysynthetic twins. Plagioclases range in size from 0.5 to 1.5 mm and make up about 50%–55% of the sample modal contents (fig. 5*f*, 5*g*). Within some plagioclases, fine opaque minerals occur. Brown and subeuhedral to euhedral amphiboles make up 20%–25% of the sample modes, and their sizes vary from 0.5 to 1 mm. Those amphiboles are darker than those from the host volcanic rocks.

Minor minerals of diorite xenoliths include clinopyroxene, mica, apatite, and opaque. Clinopyroxenes are subeuhedral and compose about 5%–9% of the mode, and their size (0.5–1 mm) is smaller than that of the clinopyroxenes from gabbroic xenoliths. The mica in some samples occurs as one of the main minerals (10%–12% of the modal content), but in most of the samples, it could be considered a minor mineral constituent (6%–9%). Micas are euhedral and semieuhedral and range in size from 0.5 to 1.5 mm. Similar to gabbroic xenoliths, some samples of diorite xenoliths display some clinopyroxenes rimmed by amphibole and some plagioclases decomposed into clay minerals, whereas most of their amphiboles are decomposed into opaque minerals (fig. 5*h*).

Analytic Methods

The major-element compositions of minerals were acquired at the “Centre de Micro Caractérisation Raimond Castaing” (Centre National de la Recherche Scientifique, Université Toulouse III, Institut National Polytechnique de Toulouse, Institut National des Sciences Appliquées, Toulouse, France), using a Cameca SXFive electron probe microanalyzer. The operating conditions were as follows: accelerating voltage, 15 kV; beam current, 20 nA; analyzed surface, approximately $2\ \mu\text{m} \times 2\ \mu\text{m}$. The following standards were used: albite (Na), periclase (Mg), corundum (Al), sanidine (K), wollastonite (Ca, Si), pyrophanite (Mn), hematite (Fe), Cr_2O_3 (Cr), NiO (Ni), and sphalerite (Zn).

The bulk-rock major and minor elements were analyzed using an inductively coupled plasma atomic emission spectrometer (Horiba Jobin Yvon Ultima 2) at the “Pôle de Spectrométrie Océan” (PSO/Institut Universitaire Européen de la Mer [IUEM], Plouzané, France), following the protocol adapted from Cotton et al. (1995). Each powdered sample was digested in Teflon vials with HF 32 N and HNO_3 14.4 N, and the dry residue was dissolved in a H_3BO_3 solution. Whin Sill dolerite England (WSE) and Ailsa Craig granite (ACE) international standards were used as internal and external controls, respectively. The precision of measurements performed on that instrument is usually better than 1% for both SiO_2 and TiO_2 , 2% for Al_2O_3 and Fe_2O_3 , and better than 4% for the other major oxides.

Trace-element concentrations were determined using a high-resolution inductively coupled plasma mass spectrometer Thermo Finnigan Element II (PSO/IUEM), following a method adapted from Barrat et al. (2007). About 100 mg of sample was digested in Teflon vials within distilled HF 32 N and HNO_3 14.4 N. The dry residue was dissolved in HNO_3 14.4 N and then in HCl 3 N. An aliquot of the solution was diluted in HNO_3 0.3 N and spiked with Tm. Basalt from Columbia River (BCR-2) and Basalt from Essey-la-Côte (BE-N) were used as external control, and Basalt from Hawaiian Volcanic Observatory (BHVO-2) was used for internal control. The precision of measurements was better than 5% for all elements.

Samples were also analyzed for Sr and Nd radiogenic isotopes. About 200 mg of sample material was weighed and dissolved in Savillex beakers in a mixture of ultrapure distilled HF (24 N) and HNO_3 (14 N) for 4 days at 120°C on a hot plate. Sr and Nd fractions were chemically separated using the Eichrom specific resins TRU-spec, Sr-spec, and Ln-spec, following conventional column chromatography procedure (Pin and Santos Zalduegui 1997). The Sr and Nd isotope compositions were measured in static mode on a Thermo TRITON spectrometer at the PSO/IUEM. All measured Sr and Nd ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively, and $^{87}\text{Sr}/^{86}\text{Sr}$ values were normalized to the recommended value of NBS987 (0.710250). During the course of analysis, Nd standard solution La Jolla gave 0.511859 ± 2 ($n = 2$; recommended value: 0.511850) and JNdi gave 0.512117 ± 3 ($n = 1$; recommended value: 0.512100).

Results

Mineral Chemistry. *Clinopyroxene.* The composition of clinopyroxenes from gabbro and diorite

xenoliths and from host volcanic rocks plot in the field of Quad (Ca-pyroxene) in the Q-J diagram (Morimoto and Kitamura 1983; fig. 6a). More precisely, all investigated clinopyroxenes are diopsides (fig. 6b). As mentioned in Petrography, gabbro xenoliths are divided into two groups based on the petrographic characteristics and mineral chemistry. In the Ti versus Ca+Na diagram (Leterrier et al. 1982), the positions of Group 1 and Group 2 gabbro xenoliths are different (fig. 6c). The composition of clinopyroxenes could provide valuable information about the magmas from which they crystallized. The Ti versus Ca+Na diagram shows that clinopyroxenes from Group 1 gabbro xenoliths and host volcanic rocks plot within the alkaline field, whereas those from Group 2 lie on the boundary between alkaline and tholeiitic/calk-alkaline fields; clino-

pyroxenes from diorite xenoliths plot within the tholeiitic/calk-alkaline field nearby the alkaline field (fig. 6c). The positions of all investigated clinopyroxenes in the Na+Al(IV) versus Cr+2Ti+Al(VI) diagram (Schweitzer et al. 1979) indicate that, except for two samples of diorite, they crystallized under conditions of high oxygen fugacity (fig. 6d). The Mg/Mg+Fe²⁺ of clinopyroxenes from Group 1 gabbro, Group 2 gabbro, and diorite xenoliths and host volcanic rocks varies from 0.82 to 0.86, from 0.74 to 0.76, from 0.62 to 0.63, and from 0.80 to 0.83, respectively (table S1 [tables S1–S6 are available online]).

The clinopyroxenes from xenoliths do not show systematic compositional variation from core to rim (table S1); therefore, their composition could be considered homogeneous. Clinopyroxenes from

Figure 6. *a*, Composition of clinopyroxenes in the Q-J diagram ($J = 2Na$, $Q = Ca+Mg+Fe^{2+}$). *b*, Classification of clinopyroxenes (Morimoto et al. 1988). *c*, Plot of Ti versus Na+Ca for clinopyroxenes from the host volcanic rocks and gabbro and diorite xenoliths. *d*, Plot of Na+Al(IV) versus Al(VI)+2Ti+Cr for clinopyroxenes from the samples.

host volcanic rocks display cores respectively higher in Al_2O_3 , FeO(t) , and TiO_2 and lower in MgO and SiO_2 than their rims (table S1). Figure 6b illustrates the compositional difference between clinopyroxenes from the different types of investigated xenoliths and from host volcanic rocks. Clinopyroxenes from Group 1 gabbro xenoliths display higher Al_2O_3 , CaO , and TiO_2 contents and lower MgO , MnO , SiO_2 , and FeO contents than those of clinopyroxene from Group 2 gabbro xenoliths (table S1). The clinopyroxenes from diorite xenoliths differ from those of both Group 1 and Group 2 gabbro xenoliths. They are different in terms of their contents of Al_2O_3 , MgO , FeO , MnO , Na_2O , and CaO (table S1). Finally, clinopyroxenes from the three groups of xenoliths differ from those from host volcanic rocks in terms of their contents of Al_2O_3 , MgO , FeO , MnO , Na_2O , and CaO .

We also compared the clinopyroxenes from the xenoliths with those from Misho gabbros (Azimzadeh 2013), located near the study area. In both cases, the clinopyroxenes are diopsides (fig. 6b), but clinopyroxenes of gabbro and diorite xenoliths from the study area differ with those from Misho gabbros in terms of MgO , Na_2O , SiO_2 , TiO_2 , Al_2O_3 , and MnO contents; in figure 6b, their positions are not the same. Consequently, it seems that the gabbro and diorite xenoliths do not belong to the same magmatic suite as the Misho gabbros.

Mica. The naming of micas was determined according to Speer (1984) and using the Al(IV) versus $\text{Fe}^{2+}/\text{Fe}^{2+}+\text{Mg}$ diagram. In this diagram, samples from Group 1 gabbro xenoliths and host volcanic rocks plot within the biotite field, whereas samples from Group 2 gabbro and diorite xenoliths plot within the phlogopite field (fig. 7a). The $\text{Mg}/\text{Mg}+\text{Fe}^{2+}$ in micas from Group 1 gabbro, Group 2 gabbro, and diorite xenoliths and host volcanic

rocks varies from 0.79 to 0.81, from 0.59 to 0.62, from 0.77 to 0.79, and from 0.78 to 0.82, respectively (table S2). The Ti content of biotites from Group 1 gabbro xenoliths is higher than those of the two other groups of xenoliths and of host volcanic rocks (table S2).

The ternary diagram $\text{MgO-Fe+MnO-10TiO}_2$ was used to assess the origin of micas (Nachit et al. 2005). According to this diagram, micas from Group 2 gabbro xenoliths and host volcanic rocks are primary magmatic micas, whereas those from Group 1 gabbro and diorite xenoliths are of recrystallized type (fig. 7b). The biotites from Group 1 gabbro xenoliths show an increase of SiO_2 , Al_2O_3 , and MgO contents and a decrease of that of K_2O from core to rim. The micas from Group 2 gabbro and diorite xenoliths do not show this kind of systematic compositional variation from core to rim (table S2). The composition of mica from Group 1 and Group 2 gabbro xenoliths differs as shown in figure 7a. The micas from Group 1 have higher SiO_2 , MgO , and TiO_2 contents and lower FeO(t) and MnO contents than those of micas from Group 2 (table S2). The composition of micas from diorite and gabbro xenoliths differ from those of host volcanic rocks for most of the major elements, including MgO , Na_2O , SiO_2 , TiO_2 , FeO(t) , Al_2O_3 , and MnO .

Finally, the composition of micas from xenoliths from this study differ from those of Misho gabbros (Shahzaidi 2013) in terms of major elements, including MgO , Na_2O , SiO_2 , TiO_2 , FeO(t) , Al_2O_3 , and MnO . In figure 7a, their positions are different.

Amphibole. The amphiboles from all investigated xenoliths and host volcanic rocks are calcic amphiboles, as illustrated by the BNa versus $\text{BNa}+\text{BCa}$ diagram (fig. 8a; Leake et al. 1997). In the $\text{Mg}/\text{Mg}+\text{Fe}^{2+}$ versus Si diagram (Leake et al. 1997), amphiboles from Group 2 gabbro xenoliths and

Figure 7. Classification (a) and composition (b) of micas in the $10\text{TiO}_2\text{-FeO-MgO}$ diagram (symbols are as in fig. 6).

Figure 8. *a*, Composition of amphiboles in the BCa+BNa versus BNa diagram. *b*, Classification of amphiboles. *c*, Plot of Si versus K+Na+Ca for amphiboles from the investigated samples. *d*, Plot of Al(IV) versus Fe/(Fe+Mg) for amphiboles from the host volcanic rocks and gabbro and diorite xenoliths (symbols are as in fig. 6). fo_2 = oxygen fugacity.

host volcanic rocks plot on the boundary of edenite and magnesiohastingsite fields, whereas those from Group 1 gabbro and diorite xenoliths plot within the magnesiohastingsite field (fig. 8*b*). In the K+Na+Ca versus Si diagram (Sial et al. 1998), amphiboles from the gabbro and diorite xenoliths and host volcanic rocks plot within the magmatic (igneous) amphibole field (fig. 8*c*). The Mg/(Mg+Fe²⁺) ratios of amphibole from Group 1 gabbro, Group 2 gabbro, and diorite xenoliths and host volcanic rocks vary from 0.65 to 0.68, from 0.70 to 0.72, from 0.66 to 0.72, and from 0.54 to 0.58, respectively (table S3). The core and rim of amphiboles from Group 1 gabbro xenoliths do not show systematic change in composition.

The Fe²⁺/Fe²⁺+Mg ratio in amphibole of the xenoliths and host volcanic rocks is less than 0.5 (table S3). This may indicate the high oxygen fugacity of their parental melts. Accordingly, in the Al(IV) versus Fe²⁺/Fe²⁺+Mg diagram (Anderson

and Smith 1995) amphiboles from the investigated xenoliths and host volcanic rocks fall within the high oxygen fugacity field (fig. 8*d*). The amphiboles of the host volcanic rocks and xenoliths display some compositional differences (fig. 8*b*). They display different compositions for most of the major elements, including SiO₂, TiO₂, Na₂O, MgO, K₂O, FeO(t), MnO, and Al₂O₃. Furthermore, some differences occur between amphiboles from the two groups of gabbro xenoliths; for instance, amphiboles from Group 1 have higher TiO₂, Al₂O₃, CaO, and K₂O contents and lower FeO(t), SiO₂, and MnO contents than amphiboles from Group 2 (table S3).

We also compared the amphiboles of the investigated xenoliths with those of the Misho diorites (Shahzaidi 2013) and found that Misho's amphiboles are edenite (fig. 8*b*), different from those from the investigated xenoliths in terms of major elements such as SiO₂, TiO₂, Na₂O, MgO, K₂O, FeO(t), MnO, and Al₂O₃.

Figure 9. Classification of feldspars from the host volcanic rocks and gabbro and diorite xenoliths (symbols are as in fig. 6). Ab = albite; An = anorthite; Or = orthoclase.

Feldspar. According to the classification of Deer et al. (1966), the compositions of plagioclase in Group 1 gabbro, Group 2 gabbro, and diorite xenoliths are bytownite, labradorite, and andesine, respectively. In contrast, the compositions of feldspars of in host volcanic rocks are oligoclase, andesine, and sanidine (fig. 9). Zoning occurs in some volcanic plagioclase phenocrysts, with SiO_2 , Na_2O , and K_2O

contents increasing from rim to core while CaO and FeO(t) contents decrease (normal zoning; table S4). Unlike the host volcanic rocks, plagioclases from gabbro and diorite xenoliths do not show zoning (table S4).

Whole-Rock Major- and Trace-Element Chemistry. Major- and trace-element analyses of gabbro and diorite xenoliths (six samples from Group 1, four samples from Group 2 gabbro xenoliths, and four samples of diorite xenoliths) are presented in table S5. In figure 10, all samples plot within the alkaline field (Cox et al. 1979), similar to the host volcanic rocks (Khezerlou et al. 2017). In this diagram, all samples plot within the gabbro and syenodiorite fields. The $\text{Mg}/\text{Mg}+\text{Fe}^{2+}$ values of Group 1 gabbro, Group 2 gabbro, and diorite xenoliths range from 0.36 to 0.65, from 0.45 to 0.68, and from 0.43 to 0.62, respectively.

Samples of Group 1 gabbro, Group 2 gabbro, and diorite xenoliths in rare earth element (REE) diagrams (normalized to chondrite values from Boynton 1984) show light REE (LREE) enrichment over heavy REEs (HREEs) and middle REEs (fig. 11b, 11d, 11f). The LREE enrichment over HREE in Group 1 xenoliths is greater than that of Group 2 gabbro xenoliths. Accordingly, these samples are characterized by enrichment in Ba, Cs, U, Pb, Sr, Sm, and Nd and depletion in Zr, Rb, Th, and Nb. The K content in these xenoliths is different (fig. 11a, 11c, 11e). Most of the samples show slightly negative anomalies in La, Ce, and P (fig. 11a, 11c, 11e); the content of these elements in the Group 2 gabbro xenoliths is

Figure 10. Samples of gabbro and diorite xenoliths plotted in the total alkali silica diagram. The line separates subalkaline from alkaline rocks. Most of the data plot in the alkaline field (symbols are as in fig. 6).

Figure 11. Primitive mantle-normalized spidergrams (a, c, e) and chondrite-normalized rare earth element patterns (b, d, f) of gabbro and diorite xenoliths. Chondrite and primitive mantle values are from Boynton (1984) and Sun and McDonough (1989), respectively (symbols are as in fig. 6).

lower than in the Group 1 gabbro and diorite xenoliths (table S5). As shown in figure 11b, 11d, and 11f, the REE patterns of the Group 1 gabbro, Group 2 gabbro, and diorite xenoliths are not parallel. The $(\text{Ce}/\text{Yb})_N$ ratios in Group 1 gabbro, Group 2 gabbro, and diorite xenoliths are 8.8–23.5, 4.5–18.9, and 11.1–29.8, respectively. Samples of Group 1 gabbro xenoliths display negative Eu anomaly, whereas those

of Group 2 gabbro and the diorite xenoliths samples do not display an obvious Eu anomaly (fig. 11b, 11d, 11f). In most samples of gabbro and diorite xenoliths (with the exception of two samples), a negative Th anomaly is observed (fig. 11a, 11c, 11e).

Whole-Rock Sr-Nd Isotopes. Three samples of Group 1 gabbro, Group 2 gabbro, and diorite xenoliths were analyzed for Sr and Nd radiogenic isotopes. The

analyses and sampling locations are presented in table S6 and figure 12. The results show that the $^{86}\text{Sr}/^{87}\text{Sr}$ ratios in Group 1 gabbro, Group 2 gabbro, and diorite xenoliths are 0.706799, 0.705120, and 0.705177, respectively, whereas the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are 0.512601, 0.512709, and 0.512665, respectively. Samples from Group 1 gabbro xenoliths plot within enriched mantle type I (EMI) and enriched mantle type II (EMII) end-members, whereas those from Group 2 gabbro and diorite xenoliths fall within EMI and high-U/Pb mantle (HIMU) fields. To validate the results, some host volcanic rocks and pyroxenite xenoliths (Khezerlou et al. 2017) from the study area and some Misho diorites (Shahzaidi 2013) were used (table S6). As illustrated in figure 12, the host volcanic rocks fall between EMI and upper crust fields (closer to the upper crust). In addition, Group 2 pyroxenite xenolith samples fall within EMI and EMI fields near Group 1 gabbro xenoliths, but samples from Misho diorite plot far from xenoliths and host volcanic rocks and closer to the EMI end-member.

Discussion

Pressure-Temperature Conditions. Given the presence of clinopyroxene in all gabbro and diorite xenoliths, the composition of this mineral was used for thermobarometry according to the method proposed by Putirka (2008). According to this method, if K_D (the partition coefficient of Fe-Mg between

clinopyroxene and melt) is within 0.28 ± 0.08 , then we may assume that the clinopyroxene is in equilibrium with the melt. Because the K_D of the investigated xenoliths (0.25–0.29) and host volcanic rocks (0.26–0.27) falls within this range (table S1), estimates may be obtained from Putirka's (2008) equation 32d for thermometry and equation 32c for barometry. According to equation 32d, the formation temperatures of clinopyroxene from Group 1 gabbro, Group 2 gabbro, and diorite xenoliths and host volcanic rocks were estimated at 1075°–1110°C, 1140°–1170°C, 1070°–1125°C, and 1120°–1140°C, respectively (table S1). The pressure of the formation of clinopyroxene in xenoliths and host volcanic rocks is calculated by the equation 32c (Putirka 2008). Using this equation, the estimate of pressure formation of clinopyroxene from Group 1 gabbro xenoliths (0.8–0.9 GPa) is higher than that for Group 2 gabbro (0.2–0.4 GPa) and diorite xenoliths (0.1–0.6 GPa) and host volcanic rocks (0.4–0.5 GPa; table S1). The depth of crystallization of clinopyroxene from Group 1 and Group 2 gabbro xenolith samples was estimated at 24–27 and 6–12 km, respectively; clinopyroxene of diorite xenolith samples and host volcanic rocks crystallized at depth ranging from 3 to 18 and from 12 to 15 km, respectively (table S1). Because the crustal thickness in the area is about 40–45 km, as determined by geophysical studies (Dehghani and Makris 1984), gabbro and diorite magmas for the xenoliths formed within the middle crustal level.

Figure 12. $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ diagram for the whole rock of investigated samples. The fields for depleted mid-ocean ridge basalt mantle (DMM), high-U/Pb mantle (HIMU), enriched mantle type I (EMI), enriched mantle type II (EMII), and upper crust are from Zindler and Hart (1986). Low-Nd (LoNd) array is from Hart et al. (1986). Misho data are from Shahzaidi (2013), and pyroxenite xenoliths and host volcanic data are from Khezerlou et al. (2017; symbols are as in fig. 6).

Oxygen Fugacity. The differences observed in the Fe^{3+} content of clinopyroxene from the investigated samples could be ascribed to different oxidation and oxygen fugacity states of the magmas they crystallized from (Canil and Fedortchouck 2000; Aydin 2008). Schweitzer et al. (1979) stated that oxygen fugacity is directly related to the formation pressure of clinopyroxene; the higher the $\text{Na}+\text{Al}^{\text{IV}}/\text{Cr}+2\text{Ti}+\text{Al}^{\text{VI}}$ ratio, the higher the oxygen fugacity. Therefore, the higher $\text{Na}+\text{Al}^{\text{IV}}/\text{Cr}+2\text{Ti}+\text{Al}^{\text{VI}}$ ratios of clinopyroxenes from the host volcanic rocks and gabbro xenoliths, compared with those of diorite xenoliths, indicate their formation at higher oxygen fugacity and probably higher pressure (table S1). The $\text{Fe}^{2+}/\text{Fe}^{2+}+\text{Mg}$ ratio in amphiboles is inversely related to oxygen fugacity (Anderson and Smith 1995). The lower values of this ratio in amphiboles from the gabbro and diorite xenoliths, compared with those of host volcanic rocks, seem to indicate that they crystallized at higher oxygen fugacity (fig. 8d). The difference in Ti content of micas from xenoliths and host volcanic rocks (table S2) implies different temperature, pressure, and oxygen fugacity at formation (Arima and Edgar 1981; Henry 1981).

Characteristics of Parental Magmas and Their Sources. According to Deer et al. (1978), if $\text{Mg}/\text{Mg}+\text{Fe}^{2+}$ of clinopyroxenes is higher than 0.7, this mineral might have crystallized from a primary magma. In such a magmatic scheme, Fe-rich clinopyroxenes represent crystallization products from differentiated magmas. Considering that the $\text{Mg}/\text{Mg}+\text{Fe}^{2+}$ ratio in the investigated gabbro xenoliths and host volcanic rocks is higher than 0.7 (table S1), their clinopyroxenes could be considered as crystallization products from primary magmas. Conversely, as this ratio is lower than 0.7 in diorite xenoliths, their clinopyroxenes could be considered crystallization products from differentiated magmas.

Because the amphiboles in xenoliths and host volcanic rocks are of igneous type, it may be deduced that their magmatic sources should contain a relatively high amount of water (Otten 1984). The anorthite content of plagioclases is typically a good indicator of the water content of their parental magma (Sisson and Grove 1993). Thus, the plagioclases in the Group 1 gabbro xenoliths might have crystallized from magma with higher H_2O content than the magma from which the Group 2 gabbro and diorite xenoliths and host volcanic rocks crystallized (fig. 9).

The minerals in gabbro and diorite xenoliths do not show clear zoning (tables S1–S4). Accordingly, they could be considered homogeneous in composition, suggesting their gradual crystallization and the equilibrium state between the melts and those

two minerals during the crystallization process. The normal zoning observed in the clinopyroxene and plagioclase from the host volcanic rocks (tables S1, S4) indicates a temperature decrease of the residual melt as crystallization proceeded.

Micas in xenoliths and host volcanic rocks contain very low $\text{Al}(\text{VI})$ content, sometimes absent (table S2), suggesting a magmatic origin (Nachit et al. 2005). The positions of xenoliths and host volcanic rocks in the ternary diagram MgO , $\text{FeO}+\text{MnO}$, and 10TiO_2 also suggest this magmatic origin (fig. 7b).

The gabbro and diorite xenoliths are subalkaline to slightly alkaline in affinity, whereas host volcanic rocks are strongly alkaline (Khezerlou et al. 2017), suggesting that they do not derive from similar magmas. Differences between the compositions of minerals of the xenoliths and those of host volcanic rocks, as well as differences between $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of xenoliths and those of host volcanic rocks, are also in agreement with this statement. Differences observed for K and Ba contents in the gabbro xenoliths are related to the variation of the mica modal content (fig. 11a, 11c). A relatively high LREE/HREE ratio associated with negative Nb and Ta anomalies in xenoliths indicates that parental magmas probably formed in a subduction zone setting (Saunders et al. 1980; Gill 1981; Pearce 1983; Wilson 1989; Kelemen et al. 1993; Kuster and Harms 1998; Barth et al. 2000; Gertisser and Keller 2000; Patchett and Chase 2002; Rudnick and Gao 2003; Green 2006).

High contents of large ion lithophile elements (LILEs) and LREEs in similar rocks worldwide are explained as representative of a metasomatized mantle, so such a mantle could be considered the source of the parental magmas of the gabbro and diorite xenoliths (Turner 1996; Alther et al. 2008; Boari et al. 2009). Because the modal contents of minerals in Group 1 and Group 2 gabbro xenoliths are similar, we may relate the higher amounts of LILEs and LREEs in Group 1 samples compared with Group 2 sample gabbro xenoliths to greater enrichment of the source of the parental magma of Group 1 samples. The higher $^{86}\text{Sr}/^{87}\text{Sr}$ ratio in Group 1 gabbro xenoliths compared with those of Group 2 is also in agreement with this statement. Positive Ba, La, Cs, K, and U anomalies associated with negative Nb, Ce, and Ta anomalies may be explained by magma contamination by crustal materials (Hofmann 1997). Nb concentration remains constant during metasomatic processes related to fluids and melts linked to the subducting oceanic plate that enriched the mantle wedge in other incompatible elements such as Ba, La, Cs, and U, leading to the negative Nb anomaly (Chappell 1999). Negative anomalies of Nb and Zr in

the investigated xenoliths may suggest the involvement of continental crust materials in their genesis. In addition, their positive Pb anomaly in xenoliths indicates a mantle wedge metasomatism through fluids from subducting oceanic plate (Atherton and Ghani 2002).

In investigated xenolith samples (except diorite xenoliths), a negative Th anomaly is observed (fig. 11a, 11c, 11e). Because Th is a very incompatible element that prefers magma during partial melting, it must not show such a negative anomaly. But this element has low mobility in arc settings because of its very low solubility in subduction zone fluids (Pearce and Peate 1995). Consequently, the observed negative Th anomaly is probably the imprint of fluids released from a subducting oceanic plate. The positive Th anomaly observed in one sample of diorite may suggest that this element might have been added to the magmatic source of this sample through a melt released from the subducting plate.

As shown in table S5, the $(\text{Ce/Pb})_N$ ratio in the investigated xenoliths is less than 18; this low ratio implies the presence of crustal materials in the magmatic sources of these rocks (Hofmann et al. 1986; Yang et al. 2005). Previous studies of volcanic rocks and pyroxenite xenoliths from the study area also suggest a metasomatized mantle source, leading to the enriched characteristics of the parental magma (Ahmadzadeh et al. 2010; Khezerlou et al. 2017). A negative Eu anomaly in Group 1 gabbro xenoliths suggests Ca-plagioclase fractionation from the parental magma during its upward migration and stay within the crust (fig. 11b, 11d, 11f; Wu et al. 2003). The lack of Eu anomaly in Group 2 and diorite xenoliths shows that plagioclase is not present as a residual mineral in the source and that melting of this source occurred under oxidizing conditions (Wilson 1989; Martin 1999). The different shape of REE patterns and composition of clinopyroxene, amphibole, mica, and plagioclase, as well as $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ values, from Group 1 gabbro, Group 2 gabbro, and diorite xenoliths are evidence of different magmatic sources.

Petrogenesis. The Ti content in clinopyroxene is a good indicator of the degree of depletion of mantle sources (Pearce and Norry 1979). In this regard, the difference in Ti concentration of clinopyroxene from gabbro and diorite xenoliths and their host volcanic rocks (table S5) may be attributed to variable depletion and degree of partial melting of their respective magmatic sources.

The chondrite-normalized HREE slope can give valuable information on the magma source. This slope for melts with a spinel peridotite mantle source is almost flat and smooth (i.e., $[\text{Dy/Yb}]_N < 1.06$),

whereas the slope is higher for magmas with a garnet peridotite mantle source (i.e., $[\text{Dy/Yb}]_N > 1.06$; Blundy et al. 1998; Shehata and Theodoros 2011). Thus, the $[\text{Dy/Yb}]_N$ ratio higher than 1.06 for the Group 1 gabbro xenoliths (table S5) may imply occurrence of garnet in the mantle source, whereas this ratio in the Group 2 gabbro (1.3–2) and diorite (1.4–1.9) xenoliths indicates occurrence of garnet and a little bit of spinel in the mantle source of their parental magmas.

To identify the characteristics of the mantle sources of gabbro and diorite xenoliths, the Sm/Yb versus La/Sm diagram was used (fig. 13). In this diagram, the depleted mid-ocean ridge basalt mantle (DMM) data are from McKenzie and O'Nions (1991) and the primary mantle (PM) data are from Sun and McDonough (1989). The Group 1 gabbro xenolith samples plot on the garnet lherzolite curve, and the Group 2 gabbro and diorite xenoliths samples plot between the garnet lherzolite and spinel-garnet lherzolite curves. According to figure 13, xenoliths plot away from the DMM and the PM, and this suggests that the xenolith parental magmas derived from an enriched lithospheric mantle. The presence of garnet with a little bit of spinel in the source of the Group 2 gabbro and diorite xenoliths and the dispersion of xenolith samples indicate the difference in the degree of partial melting and enrichment of their sources.

Generation of the Xenoliths and Volcanic Rocks

The origin of intraplate alkaline rocks is mainly related to mantle plume activity (Wilson 1989). However, the alkaline rocks from the study area have geochemical characteristics such as negative Nb, Ta, and Ti anomalies not in agreement with a mantle plume origin. As mentioned in "Geological Setting," the study area is located in the Uromieh-Dokhtar magmatic zone. Because the Uromieh-Dokhtar zone is oriented in the NW-SE direction and parallel to the Zagros Orogenic Belt, many researchers attribute the magmatism of this zone to the subduction of the Neotethys oceanic plate under the Iranian continental crust (e.g., Hassanzadeh 1993; Mohajjel et al. 2003) during the middle Cretaceous (Ghasemi and Talbot 2006), with total closure in the Upper Cretaceous (Nicolas 1989; Searle and Cox 1999). The subduction of the Neotethys led to metasomatism of the mantle wedge located above the subducting plate. Based on the geochemical and isotopic results such as high LREE/HREE ratio, negative Nb and Ta anomalies, and $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ values in xenoliths, it seems that the continental lithospheric mantle of the study area was metasomatized during the subduction of the Neotethys. As the subduction continued, the subducting oceanic plate broke (slab

Figure 13. Plots of Sm/Yb versus La/Sm for gabbro and diorite xenoliths from the NW Marand (Aldanmaz et al. 2000; symbols are as in fig. 6). DMM = depleted MORB mantle; E-MORB = enriched mid-ocean ridge basalt; gt = garnet; N-MORB = normal mid-ocean ridge basalt; PM = primary mantle; sp = spinel.

break-off). The uplift of hot asthenosphere through the window led to melting of the metasomatized lithospheric mantle (Kheirkhah et al. 2009). The formation of upper Miocene adakites in the Uromieh-Dokhtar magmatic zone is also attributed to the break-off of the subducting plate (Jahangiri 2007; Omrani 2008; Hassanzadeh et al. 2009). Postcollision tension processes have activated faults in the study area. As shown in figure 2, the studied alkaline volcanic rocks are commonly located along the main faults; it seems that the movement of the main faults (especially the North Tabriz, North Misho, and Tasuj faults) provided a way for the different parental magmas to attain their respective crystallization levels within the crust. The xenoliths formed at different depths within the crust and transported to the surface during subsequent magmatic activity.

Conclusions

Gabbro and diorite xenoliths of subalkaline to slightly alkaline affinity occur within alkaline volcanic rocks of the Uromieh-Dokhtar magmatic belt located along the main faults (especially North Tabriz, North Misho, and Tasuj faults) from the northwest of Marand. The chemical composition of the minerals and the values of $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the xenoliths and volcanic rocks are not the same, suggesting that they do not have the same parental magmas. Negative anomalies of Nb and Zr

may be explained by the addition of crustal material to the sources of the xenolith parental magmas. The negative Th anomaly in xenoliths indicates that fluids extracted from the subducted Neotethys played a major role in mantle enrichment of LREEs and LILEs. The differences in LREE and LILE values and $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in the investigated samples also suggested different enrichment levels of the mantle source of the xenolith magmas. Comparison of LILE and LREE contents and $^{86}\text{Sr}/^{87}\text{Sr}$ ratios between Group 1 and Group 2 gabbro xenoliths show that the enrichment of the source of Group 1 is greater than that of Group 2. Investigation of the chemical composition of clinopyroxene, amphibole, and mica from xenoliths and volcanic rocks shows that their parental magmas formed at high oxygen fugacity—higher for xenoliths than for host volcanic rocks. The xenolith parental magmas originated from an enriched mantle and crystallized within the crust at different depths. Group 1 gabbro xenoliths (24–27 km) crystallized at greater depth than Group 2 gabbro (6–12 km) and diorite xenoliths (3–18 km). The slab break-off and the movement of the regional faults caused the formation and intrusion within the crust of the parental magmas of the investigated xenoliths. Because the gabbros and diorites from Misho were located near the study area, it seemed plausible that the investigated gabbro and diorite xenoliths could be related to them. However, the chemical composition of the minerals and the $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$

ratios revealed that the investigated xenoliths are not linked to the Misho diorites and gabbros.

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