Magma storage and plumbing of adakite-type post-ophiolite intrusions in the Sabzevar ophiolitic zone, NE Iran

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Abstract

Subduction-related adakite-type intrusive rocks emplaced into the late Cretaceous-Paleocene Sabzevar ophiolite zone, NE Iran, range from Mg-andesite to rhyodacite in composition. Here we investigate the magma supply system to these subvolcanic intrusive rocks by applying thermobarometric mineral and mineral-melt equilibrium models, including amphibole thermobarometry, plagioclase-melt thermobarometry and clinopyroxene-melt barometry. Based on the results of these thermobarometric models, plagioclase crystallized dominantly at pressures of ~350 (130 to 468) MPa, while amphiboles record both low pressures (~300 MPa) and very high pressures (>700 MPa) of crystallization. The latter is supported by the calculated pressures for clinopyroxene crystallization (550 to 730 MPa). The association of amphibole with clinopyroxene and no plagioclase in the most primitive samples (Mg-andesites) is consistent with amphibole fractionation from very hydrous magmas at deep crustal levels of the plumbing system, which may have been a key process to intensify adakite-type affinities in this rock suite. Barometry, combined with frequent disequilibrium

features, such as oscillatory-zoned and sieve-textured plagioclase crystals with An-rich overgrowths in more evolved samples, imply final magma differentiation occurred in an open upper crustal magma system that developed progressively stronger compositional modifications during high-level magma storage.

Keywords: Adakite, Sabzevar ophiolite, Iran, mineral- melt barometry, open magma system.

1 Introduction

The Sabzevar ophiolitic zone (SOZ) in northeast Iran is part of the eastern Tethyan ophiolite belt and represents a remnant of the Cretaceous Tethyan ocean lithosphere which was obducted in late Cretaceous to early Paleocene time (Shojaat et al., 2003). The ophiolite belt contains ultramafic rocks (harzburgite, dunite and lherzolite), small masses of gabbro, and a thick sequence of submarine basaltic lavas (Shojaat et al., 2003; Khalatbari Jafari et al., 2013). The ophiolite is intruded by widespread subvolcanic stocks and dykes of intermediate to felsic compositions and the main purpose of this study is to investigate the compositional spectrum and ascent history of these post-ophiolite subvolcanic rocks. Although the petrology and tectonomagmatic evolution of the Sabzevar ophiolite belt was investigated by a number of studies previously (Lensch et al., 1977, 1980; Noghreyan, 1982; Majidi, 1999; Shojaat et al., 2003; Khalatbari Jafari et al., 2013; Shafaii Moghadam et al. 2013 and etc), the detailed evolution of the post-ophiolite subvolcanic rocks has not been thoroughly investigated. The exceptions include a number of studies such as Spies et al. (1983), who reported on the volcanic rocks in the region between Sabzevar and Quchan (North Iran) and highlighted the presence of subalkaline rocks that range from 41 ± 2.1 to 2.7 ± 0.2 Ma in age, and that become younger northward, reaching Late Miocene to Late Pliocene ages in the Quchan. Ghasemi et al. (2010) and Shabanian et al. (2012), working on Sarakhor adakitic domes

(Southern of Quchan), suggested post-collisional melting of an oceanic slab and/or mafic the mafic lower crust as a source for the post-ophiolite intrusive suite, possibly triggered by an asthenospheric rise after slab break-off. Mohamadi Gorani et al. (2013) suggested adakitic signatures of some felsic domes in the Southwestern part of the Sabzevar ophiolite belt and attributed their origin to partial melting of the subducted oceanic slab. Jamshidi et al. (2014), studying the same subvolcanic domes in the north and northwest of Sabzevar zone, interpreted this adakite magmatism as partial melts derived from a garnet-amphibolite source that originated from metamorphism of the Sabzevar Neotethyan subducted oceanic slab underneath the southern edge of the eastern Alborz zone. The ascending magma was then subjected to crustal differentiation processes such as assimilation and fractional crystallisation during ascent and emplacement into the ophiloitic belt (Jamshidi et al., 2014). Rossetti et al. (2014) reported similar low-Mg# adakite intrusive rocks in the Soltanabad area (northeast of Sabzevar zone), and suggest that these rocks formed from high-pressure wet amphibolite melting of the subducting oceanic slab, and later amphibole fractionation.

Here, we present new major and trace element data on whole rocks plus major element data from the main mineral phases in these to improve our knowledge on the history of the post-ophiolite intrusive suites in the wider region of the Sabzevar ophiolite belt. We attempt to compare post-ophiolite felsic rocks in the southern Sabzevar belt with dominantly intermediate ones in the North and aim to test if differences between the post-ophiolite subvolcanic rocks of the two sectors exist. Finally, we hope to establish the P-T conditions of amphibole and plagioclase crystallization through mineral-melt equilibrium thermobarometry to help reconstruct the former plumbing system to these rocks.

Amphibole occurs as essential rock forming mineral in a wide variety of igneous and metamorphic rocks and is especially abundant in calc-alkaline igneous rocks. Experimental studies have synthesized amphibole over a pressure range of up to 2300 MPa and 400° to

1,150°C in calc-alkaline compositions (Blundy and Holland, 1990) and amphibole has therefore considerable potential as an indicator of crystallization conditions, both as geothermometer and geobarometer (Blundy and Holland, 1990; Ridolfi, 2010).

The composition and growth morphology of igneous feldspar usually reflects progressive changes in crystallization conditions that give a reliable record of crystallization dynamics of an associated melt and its thermal and compositional history (e.g. Troll and Schmincke, 2002; Slaby and Götze, 2004). In the studied rocks, amphibole and plagioclase are found most widely in the northern suite of intermediate rocks and we employed the thermobarometers of Ridolfi and Renzulli (2012) for selected amphiboles to determine crystallization and storage conditions for these rocks. Plagioclase-liquid thermobarometry (Putirka, 2008) and clinopyroxene-melt barometry (Putirka, 2008) were employed in order to complement and test the results from amphibole thermobarometry.

2 Geological setting

The Sabzevar ophiolite zone is ~200 km long and ~10 km wide and represents a tectonically dismembered ophiolite complex, located along the northern boundary of the Central Iranian microcontinent (Shojaat et al., 2003). Plate tectonic reconstructions suggest that the Sabzevar ophiolite was part of the Tethyan Ocean that formed during Cretaceous rifting from a narrow but deep seaway, separating the Central Iranian microcontinent from the Eurasian plate. The ophiolite was emplaced during an episode of northeast dipping subduction (i.e. closure) of the Tethyan seaway (the Sabzevar Ocean) in Upper Cretaceous- early Paleocene time (Shojaat et al., 2003). The emplacement of the Sabzevar ophiolitic belt was followed by lasting post-ophiolite volcanism from at least the Eocene (~40 Ma) to the end of the Pliocene (~2.3 Ma) (Lensch et al., 1977; Spies et al., 1983; Shojaat et al., 2003). This Eocene volcanism is expressed in central and in northern Iran (Berberian and King, 1981; Berberian et al., 1982;

Bina et al., 1986; Stampfli and Borel, 2002; Agard et al., 2005; Shabanian et.al, 2012) and Eocene volcano-sedimentary rocks cover extensive areas in the Sabzevar region. Since the Oligocene, volcanism has been intermittent (Jahangiri, 2007) and, for ~10 Ma focussed predominantly on the Turkish-Iranian plateau (e.g., Berberian and Berberian, 1981; Keskin et al., 1998; Azizi and Moinevaziri, 2009). The Oligo-Miocene sedimentary rocks are composed of marine flysch-type facies in their lower parts, while younger facies are of continental origin and contain, in some parts, extremely high percentages of volcanic detritus. These sedimentary rocks are tectonically imbricated, often steeply dipping and penetrated by Oligocene alkali-olivine-basalt magmatism (Ghasemi et al., 2011). Following the Eocene volcanic and volcano-sedimentary and the Oligocene volcanic activity, magmatism recommenced in the region and continued to the end of the Pliocene. In the Sabzevar belt, characteristically, post-ophiolite subvolcanic rocks occur in the north and in the south of the ophiolite zone (Fig. 1) and magmatism comprised intermediate rocks in the northern part, while felsic rocks dominate the southern part (Fig. 2). Fragments of host rocks, including serpentinized harzburgites and tuffaceous siltstone are found in some intrusions as xenoliths (Fig. 3 a, b). Large volumes of detritus originated from these intrusive domes, forming Pliocene-Pleistocene conglomerates (Fig. 3c). According to these geological lines of evidence and stratigraphic relationships, Salehinejad (2008) and Jamshidi et al. (2014) suggested that the studied domes eventually outcropped at the Earth’s surface around the Oligocene-Miocene Pliocene boundary. Recent integrated U-Pb zircon and ^{40}Ar/^{39}Ar white mica and amphibole geochronology on similar rock suites in the Soltanabad region by Rossetti et al. (2014) constrains the Sabzevar magmatism to the late Paleocene (at ca. 58 Ma). In the absence of a comprehensive dating study, we cannot specify the age of our post-ophiolite rocks, but note that the presence of xenoliths that look like Eocene tuffaceous siltstone, makes us consider a post-Eocene age at least for some of the subvolcanic rocks in question.

3 Method

3.1 Analytical techniques

Twenty-three fresh whole rock samples from the northern sector and 34 samples from the southern sector were selected for major and trace element analysis at AMCE analytical laboratory, Vancouver, Canada. Total abundances of the major oxides were determined by ICP-emission spectrometry following a Lithium metaborate/tetraborate fusion and dilute nitric digestion (analytical code: 4A04). Loss on ignition (LOI) is by weight difference after ignition at 1000°C. Fe_2O_3 is reported as total. We recalculated FeO and Fe_2O_3 according to the method suggested in Le Maitre (1976). Trace elements and REE concentrations were obtained by solution inductively coupled plasma-mass spectrometry (ICP-MS, Analytical code: Group 1T) for whole rocks. For further analytical details and uncertainties please see 'www.acmelab.com'. Major, trace element data for 57 post-ophiolite subvolcanic rocks are listed in Table 1.

Major element analyses on selected minerals (amphibole, plagioclase and pyroxene) were performed on Jeol JXA 8530F Hyperpobe at CEMPEG. Analytical conditions included an accelerating voltage of 15 kV, a beam current of 10nA and counting times of 10s on peaks and 5s on +/- background. For calibration of all elements a set of mineral and synthetic standards has been used. All elements were analysed using K (alpha) lines. The analysed data set consists of 197 points on amphibole, 212 on plagioclase and 29 on clinopyroxene, collected from 30 amphibole, 26 plagioclase and five complementary clinopyroxene crystals. Representative compositions of selected minerals are given in Table 2, 3 and 4.

3.2 Amphibole thermobarometry

Temperature, pressure, oxygen fugacity and H_2O_{melt} conditions were estimated from amphibole compositions using the recent thermobarometric formulation of Ridolfi and Renzulli (2012), which is able to estimate the P, T and fO_2 in a wide range of amphibole crystallization conditions. Their new single - crystal model (requiring only the amphibole compositions) allows to estimate the physico-chemical parameters at low uncertainties (T ± 23.5 °C, P ± 11.5%, H_2O_{melt} ±0.78 wt.%) for calc-alkaline and alkaline magmas up to 1,130°C and 2,200 MPa and ΔNNO values (±0.37 log units) up to 500 MPa (NNO = nickel-nickel oxide buffers). Application of this method is limited to igneous amphibole phenocrysts and cannot be applied to fluid-related (hydrothermal) amphibole veins, to microlites or quenched amphibole zones in erupted products that likely reflect variable disequilibrium conditions (Ridolfi and Renzulli, 2012).

3.3 Plagioclase- melt thermobarometry

To test and complement the results of the amphibole thermobarometry, the plagioclase-melt thermobarometer of Putirka (2008), calibrated for hydrous systems, was put to use. The standard error of estimation (SEE) for this thermobarometer is ±36 ˚C and ±247 MPa (Putirka 2008), which is less precise than the amphibole barometer employed. The most commonly used nominal melts are whole rock, groundmass and glass compositions (Putirka et al. 2003; Putirka 2008) and the respective whole rocks were applied as the nominal melt in our study.

3.3 Clinopyroxene- melt barometry

This barometry model for clinopyroxene-melt equilibria after Putirka (2008, eqn. 32c) has been applied to cross-check the result from amphibole and plagioclase-melt barometry. This model is based on Al partitioning between clinopyroxene and coexisting melt in hydrous magmatic systems. The SEE for this barometer is ± 150 MPa and the respective input melt was the corresponding whole rock composition (cf. Putirka, 2008).

3.4 Equilibrium tests

Putirka (2008) present an equilibrium test for plagioclase-melt thermobarometry based on the partitioning coefficients of the anorthite and albite components, Kd[An-Ab]. The equilibrium constant is sensitive to temperature and should be 0.10 ± 0.05 at T<1050 C° and 0.27 ± 0.11 at T>1050 C° (Putirka, 2008). The plagioclase components which are near to equilibrium with a given melt at the given pressure and temperature conditions should fall within the $Kd_{[An\text{-}Ab]}$ envelope appropriate for this melt and mineral composition.

Clinopyroxene-melt equilibrium was tested using two models. Firstly, we looked at the partition coefficient of Fe and Mg between clinopyroxene and melt. According to Putirka (2008), the acceptable range of equilibrium values for $Kd_{[Fe\text{-}Mg]}$ is 0.28 ± 0.08. We tested our available whole rocks as nominal melts. A melt that plot within the $Kd_{[Fe\text{-}Mg]} = 0.28 \pm 0.08$ boundary can thus be assumed to represent a liquid in potential equilibrium with the analyzed clinopyroxenes. All datapoints plotting within this range were further tested using the equilibrium test presented in Putirka (1999), based on the equilibrium partitioning of Na-Ca-Al. Here, predicted values of different components that crystallize from a nominal melt are compared with observed clinopyroxene components and mineral-melt equilibrium pairs require a close match between predicted and observed components to satisfy equilibrium conditions.

4 Results

4.1 Petrography

Post-ophiolite magmatism in the northern part of Sabzevar belt comprises intermediate rocks of andesite, trachyandesite, trachydacite to dacite composition. Andesite/trachyandesite rocks intrude into the ophiolite complex and younger volcanoclastic rocks and appear as grey-green dome-forming intrusions in the field, but occur as dykes also. In thin section, the andesite/trachyandesite rocks show porphyritic, acicular and fluidal textures and contain plagioclase and amphibole as the main minerals together with variable amounts of plagioclase, biotite and Fe-Ti oxide microlites in the groundmass (Figs. 4a, b). Clinopyroxene phenocrysts of mostly diopside to augite composition are only found in a high Mg-andesite samples (Fig.4 c). Notably, this rock is a mafic basaltic andesite and shows amphibole and clinopyroxene phenocrysts, but lacks plagioclase as phenocryst phase. Amphibole phenocrysts are generally star shaped or long prisms up to 1 cm and the groundmass of the dyke is composed of fine-grained hornblende, plagioclase and augite.

Euhedral to subhedral biotite is present and in more evolved trachydacites and is moderately enriched in Mg (average Mg# ≈ 0.65). Monomineralic glomerocrysts of amphibole do occur occasionally in the trachyandesite samples. Accessory minerals include apatite, Fe-Ti oxides and rare alkali feldspar. Apatite is ubiquitous, occurring as euhedral needle-like phenocrystals and as inclusions in larger phenocrysts. Fe-Ti oxides (mostly magnetite) occur as euhedral microphenocrysts and as inclusions in amphibole and biotite. Some trachyandesite samples also contains small but coarse-grained and rounded crystal clots up to 2mm in diameter (Fig. 4d) that are made up of euhedral microcrystals of plagioclase, amphibole and accessory apatite.

Dacite samples from light-grey domes and dykes show plagioclase as the volumetrically dominant phenocryst phase, which usually displays polysynthetic twinning and compositional zoning (Fig. 4e). Amphibole and biotite phenocrysts are apparent and the medium to fine-grained groundmass is usually composed of plagioclase and amphibole microcrystals with biotite, Fe-Ti oxides and apatite occurring in smaller proportions.

Felsic rock types of broadly dacite to rhyolite in composition crop out as dominantly light grey to white-coloured domes in the southern and south-western part of the Sabzevar belt. Like the dacite samples from the northern part, dacite domes in south-western part represent plagioclase phenocrysts, biotite and rarely amphibole in the fine-grained groundmass. Southern rhyolite samples show Quartz, sanidine, and plagioclase phenocrysts with a fine-grained groundmass (Fig. 4f). Although smaller quartz phenocrysts have frequently straight edges, many large crystals display embayments and round edges. Some embayments are short and close to the margin, whereas others are penetrating deep into the core of the grains. The most suitable processes to explain deep embayments in quartz is dissolution in superheated and H_2O-saturated rhyolite melt due to ascent and decompression or magma mixing (e.g. Donaldson and Henderson, 1989; Chang and Meinert, 2004). White-coloured sanidine occurs too and has a subhedral form, but is often altered. Intensely altered grains of sanidine are cloudy and turbid and were not selected for microprobe analysis.

4.2 Whole rock geochemistry

For the purpose of plotting, all oxide values have been normalized to 100% (volatile free), and iron content is given as FeO_{total}. The samples from the northern part have SiO_2 contents of 55 to 64.0 wt.% and fall into the andesite, trachyandesite, trachydacite and dacite fields of the total alkali vs. silica diagram (TAS; Fig. 5a). They have low TiO_2 (0.4-0.7 wt.%) and high Al_2O_3 contents (14.2–19.1 wt.%). K_2O contents (0.6–2.9 wt.%) are less than Na_2O (3.9-7 wt.%), belonging to the calc-alkaline (Fig. 5b) and metaluminous to peraluminous (Fig. 5c) series. Mg# (100×MgO/[MgO + FeO]) ranges from 37.5 to 75 (Fig. 5d). Strong negative correlations can be observed in Harker diagrams for TiO_2, FeO_{total}, MgO, CaO and Zr/Nb ratio vs. SiO_2 , but positive trends for Na_2O, K_2O, La/Yb, Rb/Sr and Sr/Y ratios are observed (Fig. 6), indicating magmatic crystallization and associated differentiation due to enrichment in incompatible elements. Also, The decrease of Dy/Yb (Fig. 6i) combined with an increase in Sr/Y ratios with differentiation are consistent with partition coefficient of these elements between amphibole and tonalite melts, i.e., $D_{DY}/D_{Yb} < 1$, and $D_{Sr}/D_Y < 1$ (Klein et al., 1997; Bédard, 2006; Davidson et al., 2007; Tiepolo et al., 2007 in Rossetti et al. 2014). Multi elements mantle-normalized and REE chondrite-normalized plots for these samples (Fig.7a,b), show typical subduction-related trace element patterns, including enrichments in LILE elements relative to LREEs (e.g. Ba/La), but with both groups being enriched relative to their HFSEs (e.g. Ba/Nb; La/Nb). According to these plots, the rocks are also characterized by an overall enrichments in LREEs relative to HREEs of on average $(La/Yb)_N = 7.9$, U and Zr depletions in HFSEs, and absent Eu anomaly, which are typical features of adakite-type rock compositions (e.g. Defant and Drummond, 1990). The samples also contain high concentration of Sr (avg. 460 ppm) and low concentrations of Y (5-19 ppm) and Yb (0.4-2 ppm) compared to typical arc volcanics and in the Y vs. Sr/Y diagram (Fig. 8a) most of the samples plot in the field of adakite rocks as opposed to typical arc-related calc-alkaline composition (cf. Defant and Drummond; 1990). However, a few samples plot near the intersection of the adakite and the arc-related fields, while a plot of Sr against $CaO+Na_2O$ (after Castillo; 2011), shows most of our samples in the high silica adakite field (Fig. 8b).

Notably, the samples of the southern part have higher silica contents than the northern suite and range from 65 to 73.6 wt.% in SiO_2, falling into the dacite and rhyolite field of the TAS diagram (Fig. 5a). They have high alkali contents with K_2O contents of 1.3 to 4.2 wt.% and Na_2O of 3.5 to 7.1 wt.% resulting in Na_2O/K_2O ratios of 2.2 to 3.7 and the rocks classify as calc-alkaline to high-K calc-alkaline and peraluminous (Fig. 5 b and c). The Al_2O_3 contents range from 15 to 21 wt.% and the calculated Mg# numbers from 25 to 68 (Fig. 5d). Sr concentrations vary from 208 to 893 ppm, with low Y (2.2-13 ppm) and Yb (0.5-1.5 ppm) contents similar to the range observed in typical adakites (Defant and Drummond, 1990). The adakite-like geochemical features are further underlined by their fractionated REE patterns with a mean $(La/Yb)_N$ of 15.2 (Fig. 7c and d), depletions in Nb, Ta and Ti, enrichments in U and K, a mildly positive to absent Eu anomaly, and high Sr/Y and La/Yb but low Y and Yb. On the Harker diagrams (Fig. 6), a significant correlation between major and incompatible element ratios vs. SiO_2 between the southern and northern intermediate samples could imply that the magmas are connected by crustal differentiation processes, such as fractional crystallization or AFC, likely involving significant amphibole fractionation.

4.3 Mineral chemistry

4.3.1 Amphibole

Amphibole phenocrysts are found as euhedral crystals with normal zoning in the trachyandesite and dacite samples and mostly as acicular or star-shape crystals in andesite dykes from the northern part of the study area. Selected amphiboles were analysed for major element compositions. Amphibole stoichiometry and nomenclature follows the recommendations by the International Mineralogical Association (IMA) outlined in Leake et al. (1997), where the Fe^{+3}/Fe^{+2} ratio is determined by charge balance after normalization of the major element analyses on the basis 23 oxygens to 13 cations. This scheme is based on the general chemical formula $A_{0-1}B_2C_5{}^{V1}T^{1V}{}_8O_{22}(OH; F; Cl)_2$. Because the water and halogen contents of the amphiboles are unknown, the amphibole formula is calculated to 23(O). Amphiboles are classified through the number of atoms per formula unit (apfu) of Ca and Na in the B site, (i.e. $(Ca+Na)_B$) and amphiboles from the northern subvolcanic suite plot in the calcic amphibole field. These amphiboles have Si between 6 and 7 values per formula, characteristic of igneous amphiboles (Leake, 1978). The compositions of the amphiboles from the andesite, trachyandesite, trachydacite and dacite samples are shown in the $Mg/(Mg+Fe^{2+})$ vs. Si classification diagram (Fig. 9). The amphiboles from the andesite samples are Mg-hastingsites to tschermakites, and those in the trachyandesite and trachydacite samples are tschermakite, while amphiboles from dacites are mostly magnesio-hornblendes (Fig. 9c). Mg-hastingsite can be separated from pargasite by values of Al^{VI} and Fe^{+3}cations per formula; Al^{VI} values in hastingsite are less than Fe^{+3} (Fig. 9b). A plot of six-fold Al (Al^{VI}, C site) vs. four-fold Al (Al^{IV}, T site) for amphiboles from all rock suites (Fig. 9d) indicates that Al preferentially resides in the tetrahedral site, which, together with the negative correlation between $(Na+K)^{A}$ vs. Si p.f.u (Fig. 9e), implies the dominance of edenite-type ($Na,K_{A}+Al^{IV}=Si^{IV}$) substitution for the presented amphiboles (cf. Murphy et al, 2012).

Although, the amphiboles are unzoned in the andesites, simple and oscillatory zoning is present in amphiboles from the trachyandesites and trachydacites. The simple-zoned amphibole phenocrysts are characterized by broad cores that commonly host small subhedral inclusions of Fe-Ti oxides (not shown). The cores of the simple-zoned phenocrysts are tschermakite and show a progressive rimward increase in Mg that is coupled with decrease in ^{IV}Al and $(Na+K)^{A}$. The oscillatory zoned amphibole phenocrysts in the dacite and trachydacite samples are illustrated by alternating dark and bright euhedral zones in back-scattered images (Fig. 10a). This zoning is characterized by a tschermakitic core that also commonly hosts small inclusions of Fe-Ti oxides and apatite. The cores of these minerals shows a rimward decrease in Mg (from 3.44 in the core to 2.8 cations p.f.u. in the rim) and an increase in Al^{iv} (from 1.59 to 1.63 cation p.f.u) and $(Na + K)^{A}$ (from 0.30 to 0.41). The cores are overgrown by a layer of dark Mg-hornblende and show significantly lower Al^{iv} (1.36) and $(Na + K)^{A}$ (0.28) compared to the cores. Interval changes in this zone are marked by decreases in Mg (from 3.17 to 2.62 cations p.f.u) and increases in Al^{iv} (from 0.26 to 0.46 p.f.u) and (Na

+ K)^A (0.17-0.47) toward the rim. The Mg-hornblende zones then grade into outer rims with ferrian-tschermakite composition, reflecting normal zoning with a decreasing trend in Fe^{+2} (from 1to 0.41 p.f.u), Al^{iv} (1.77 to 0.59 p.f.u) and (Na + K)^A (from 0.43 to 0.33), while a very mild increase in Mg is observed (from 3 to 3.14 cation p.f.u) (Fig. 10b, c).

4.3.2 Plagioclase

Plagioclase occurs abundantly together with amphibole in the northern subvolcanic samples as individual phenocrysts, microphenocrysts, and microlites. Although plagioclase crystals commonly range between 0.5 and 3 mm, large phenocrysts up to 7 mm in length do occur in some dacite samples from the northern part. The southern rhyolite suite comprises plagioclase as phenocrysts and microlites. The composition of plagioclase phenocrysts in the andesite and trachyandesite rocks ranges from ($An_{53}Ab_{46}Or_1$) to ($An_{70}Ab_{29}Or_1$) and ($An_{36}Ab_{62}Or_2$) to ($An_{0.56}Ab_{0.43}Or_{0.00}$), respectively. Plagioclase is the volumetrically dominant phenocryst phase in the trachydacite ($An_{31}Ab_{67}Or_2$ to $An_{49}Ab_{51}Or_1$) and dacite ($An_{40}Ab_{58}Or_2$ to $An_{0.24}Ab_{0.73}Or_{0.03}$) samples, but is rare in the rhyolites ($An_{15}Ab_{80}Or_5$ to $An_8Ab_{88}Or_4$). Despite some homogeneous plagioclase phenocrysts that occur in the andesites, most of the samples in the northern part exhibit plagioclase with complex zoning of variable types. Plagioclase microlites, in turn, are usually unzoned and have homogeneous compositions or minor normal zoning. Fig.11 a shows the compositional spectrum of all plagioclase crystals analysed in the feldspars composition triangle. Notably, plagioclase phenocrysts with pronounced sieved textures are preferentially found in dacite rocks (Fig.11 b, c). Two types of primary zoning in plagioclase are dominant and include normal and oscillatory (complex) types:

Normal zoning in plagioclase implies a composition change from calcic to less calcic with progressive growth (Bottinga et al., 1996; L'Heureux, 1997). This type of zoning is seen in a fraction of plagioclase phenocrysts and most microcrystals. It lacks internal dissolution surfaces and the zoned phenocrysts range in composition from An_{50} to An_{39} and likely reflect periods of stable growth from progressively evolving melts that became depleted in Ca.

Oscillatory zoning is also present in a fraction of feldspars and consists of alternation of thin growth zones with variable An contents (cf. Allegre et al., 1981; Bottinga et al., 1996; L'Heureux, 1997). In our samples, neighbouring zones show abrupt increase in An (up to 15 mol %) and phenocrysts have typically euhedral to subhedral shapes. Oscillatory types appear particularly in trachydacite, dacite and in one of the rhyolite samples. Characteristically, the cores are marked by a slight increase in X_{An} (e.g. from An_{29} to An_{39}) followed by several zones with markedly higher anorthite contents (An_{46-55}) (Fig. 11 b, d). These An-rich compositions are then usually followed by steep normal zoning. Reverse zoned outer margins (~100 μm) often follow, and are characterized by an overall increase in X_{An} (e.g. from An_{39} to An_{44}). Simple resorption has taken place at some of the internal boundaries, as marked by wavy truncated internal surfaces.

Sieved texture plagioclase phenocrysts are frequently found in trachydacites and dacites and generally show resorbed cores and oscillatory-zoned overgrowth rims (Fig. 11 c, e). Broad patchy cores are usually albite (An_{24-34}) and are mantled by more An-rich (An_{36-47}) overgrowth rims. This increase in X_{An} is accompanied by an increase in Fe and usually follows a partial dissolution event, thus being symptomatic of magma mixing (cf. Troll and Schmincke 2002).

4.3.3 Clinopyroxene

Clinopyroxene crystals are mostly euhedral (Mg # = 0.8, in average) and are only found in the high Mg-andesite sample together with Mg-hastingsite. The mineral chemistry of clinopyroxene is very homogenous and all the analysed clinopyroxene plot in the diopside field (Fig. 12), where clinopyroxene components are calculated on the basis of 6 cations. They exhibit no clear petrographic indicators of being out of equilibrium with andesitic host rock, as they are commonly euhedral and lack strong compositional zoning. Representative clinopyroxene compositions are provided in Table (4).

4.4 Temperature and pressure from amphibole thermobarometry

High-Mg hastingsites from the andesite samples yielded temperatures of 908 to 992°C (±23.5), while tchermakitic amphiboles from the trachyandesite and trachydacite samples range from 868 to 960˚C and from 858 to 946˚C, respectively (Fig. 13 a). Magnesian hornblendes from the dacite samples range from 836 to 873˚C and estimated temperatures for amphiboles in the clots range between 874 and 974˚C.

Amphiboles from andesites yield pressures of on average 556 MPa (820-349) (Fig. 13 a). The highest crystallization pressures (820-700 MPa) correspond to the Mg-hastingsites from the basaltic andesite. Average pressure estimates of ~370 MPa (714-246) for tschermakites in trachyandesites and ~320 MPa (388-268) in trachydacites are derived. Mg-hornblendes in dacites yield pressure of ~190 MPa (226-145), while tchermakites from crystal clots gave an average pressures of ~395 MPa (780-217). To reconstruct the plumbing system, we used the density of the ophiolite complex (3.0 g/cm^3) and applied this value to the sub-ophiolite units also (see Discussion).

The oxygen fugacity of a crystallising magma is reflected in the Mg content of resulting amphiboles and values of log ($f O_2$) for Mg-hastingsites in andesites range from -10 to -8.9 (Fig. 13 b). Values of log ($f O_2$) for tschermakites from trachyandesites and trachydacites range from -11 to -9.5 and -11.0 to -9.8, respectively. Mg-hornblendes of dacites show log ($f O_2$) from -12.1 to -10.9 (Fig. 13 b). Tschermakites from crystal clots have a value of log ($f O_2$) similar to trachyandesitic samples ranging between -10.4 to -9.3 (Fig.13 b). Oxygen fugacity generally increases from high P-T magnesiohastingsite (avg. 700 MPa, 990°C) to lower P-T magnesiohornblende (avg. 190 MPa, 840°C), consistent with the compositional variations known for calc-alkaline magmas from experiments (Gill, 1981; Martel et al., 1999; Müntener et al., 2001; Behrens and Gaillard, 2006).

A diagram of H_2O_{melt}-T of the amphibole suite is given in Fig. 13 c, where the stability field of experimental amphiboles is outlined by a dotted curve. The experimental amphiboles comprise crystals with Al# ≤ 0.21 that were synthesized in equilibrium with melts overlapping the main Al_2O_3 vs. SiO_2 pattern of extrusive rocks and glasses (cf. Ridolfi et al. 2010). All studied amphibole samples fall within the range of the experimentally determined equilibrium amphiboles with low-Al# (<0.21) (Ridolfi et al., 2010). This Al-value is, in turn, in equilibrium with H_2O_{melt} values of 3.7 to 8.2 wt.% (± 0.5) (cf. Ridolfi and Renzalli, 2012), which agrees with the H_2O values known in calc-alkaline magmatic suites (e.g. Gill, 1981; Martel et al. 1999). The H_2O_{melt} contents obtained from Mg-hastingsites range between 5 and 8.3 wt.% (±0.5), while tschermakite and Mg-hornblende yield 4.2 to 10.3 and 5 to 7 wt.% (± 0.5) melt H_2O, respectively. Tschermakitic clots suggest corresponding H_2O_{melt} values between 5.5 to 9.3 wt.% (± 0.5).

4.5 Pressure estimate from Plagioclase- melt thermobarometry

To test the results of the amphibole thermobarometry, the plagioclase thermobarometer of Putirka (2008), calibrated for hydrous systems, was applied. Standard error of estimation (SEE) for the plagioclase-melt thermobarometer is ±36 ˚C and ±247 MPa (Putirka, 2008). The Kd [Ab–An] equilibrium test shows that $An_{50–70}$ is in equilibrium with the whole rock composition of the andesites and these plagioclase-melt pairs record pressures between ~470 and 335 MPa (Fig. 14 a, b). The lower-anorthite plagioclase (An_{50-30} and An_{30-20}) shows disequilibrium with the andesite nominal melt, but yields equilibrium with the available trachydacite and dacite compositions, respectively. Plagioclases with An_{50-30} record pressures between ~435 and 190 MPa and plagioclase with An_{30-20} records between ~390 and 130 MPa (Fig. 14 a, b). Plagioclase in the southern rhyolites (An_{20-10}) yield pressures between ~275 and 110 MPa. The results of the plagioclase-melt thermobarometer record overall mid- to shallow crustal magma storage (468-110 MPa), thus broadly overlapping with the amphibole thermobarometery results.

4.6 Pressure estimate from clinopyroxene- melt barometry

Following equilibrium testing between clinopyroxene and available mafic whole rocks as nominal melt (see Methods), pressure estimates were determined. The selected clinopyroxenes (sample N-1, Fig. 15 a, b) satisfy equilibrium conditions with the analysed andesite whole rock and using clinopyroxene-melt barometry after Putirka (2008, eqn 32c), average pressures of ~600 MPa (730 to 510) are retrieved (Fig. 15 c). These results are higher than the results from plagioclase-melt barometry, but overlap with pressures of the deeper fraction of amphiboles in these andesites (see above; Fig. 13 a).

5 Discussion

The Cretaceous to Paleocene Sabzevar ophiolitic zone in present-day Iran formed during northward subduction of the Tethyan oceanic crust beneath the eastern Alborz (Binalud) plate. Post-ophiolite subvolcanic rocks are widely distributed in the northern and southern parts of this zone and show characteristic subduction-related trace-elements signatures, such as negative Nb, P and Ti anomalies, coupled with enriched LILE elements. The northern post-ophiolite magmatic rocks show a broad compositional range, from andesite to dacite, while the southern rocks are exclusively rhyolite and rhyodacite in composition (Fig. 5a). All these post- ophiolite rocks fall within the adakite field on commonly used discrimination diagrams, such as Sr/Y vs. Y (Fig. 8a). Some of their adakitic affinities include the relative enrichment of Sr, depletion of Yb and Y and the negligible or absent Eu anomalies (Fig. 6k). Adakites have been originally considered as products from partial melting of subducting metabasaltic source materials (Defant and Drummond, 1990), while later work considered adakites to potentially form from lower crustal melting also (e.g. Castillo, 2012).

To help constrain the way adakites form, we first discuss the magma composition and plumbing of the northern adakites suite. These illustrate negative correlations of TiO_2, FeO_{total}, MgO, CaO vs. SiO_2 in the binary diagrams (Fig. 6), indicating crystallization-controlled differentiation trends. Lower contents of MREEs (e.g. Tb= 0.1 ppm) and Y (3.5-8 ppm) as well as decreasing in Dy/Yb with differentiation (increasing SiO_2) and strong increasing in Zr/Sm (73-89 ppm), Zr (88-186 ppm) in more evolved rocks (e.g. dacites) suggest extensive fractionation of amphibole (e.g. Davidson et al.; 2007), in agreement with its occurrence as phenocrysts in andesite to dacite samples. Prolonged amphibole fractionation is also supported by the presence of amphibole and clinopyroxene phenocrysts in the most primitive Mg-andesite that simultaneously lacks plagioclase as a phenocryst phase. The compositional variations of the amphibole reflect changes in T, P, pH_2O, fO_2 and melt composition (e.g. Rooney et al., 2010) and intense fractionation of amphibole will increase the Sr/Y ratios and decrease the Y concentrations in the resulting melt (Foley et al., 2013), driving compositions to that of typical adakite magmas. Crystal clots composed of amphibole have been suggested to represent clusters of phenocrysts (Garcia & Jacobson, 1979), the products of amphibole breakdown (Stewart, 1975), cooler wall-rock material (Humphreys et al., 2009), disrupted cumulates or crystal mush zones (e.g. Seeman, 2000; Chiaradia et al., 2011), or xenoliths (Chiaradia et al., 2009). Our investigation shows that the mineral compositions and the crystallization conditions (T, P, and fO_2) of the crystal clots overlap with the phenocryst assemblage of the host-rock, implying that the clots in the present system are associated with cooler wall-rock crystallisation.

The core to rim variations in the simple-zoned amphibole phenocrysts shows an overall rimward decrease in Al^{iv}, $(Na+K)^A$ and Fe^{+2} accompanied by an increase in X_{Mg}. This pattern can be explained by gradual cooling during crystallization (e.g. Humphreys et al., 2006) where composition of the amphibole has likely been modified in response to changing melt composition controlled by plagioclase crystallization (cf. Holland and Blundy, 1994). For example, slight increases in melt-H_2O result in dissolution and crystallization of An-rich compositions (Housh and Luhr, 1991). Although normal zoned plagioclases with small-amplitude variation are probably a result of minor local disequilibrium or small-scale pressure and temperature changes (Pearce and Kolisnik, 1990), the occurrence of disequilibrium phenocryst composition and textures such as oscillatory zoned amphibole and plagioclase and sieved textured plagioclase cannot be explained by simple closed system fractionation alone.

Oscillatory and especially complex zoning in plagioclase is often marked by resorption surfaces presented by rounded or wavy truncated interfaces (e.g. Fig. 11 b) which is attributed to large-scale changes in temperature, pressure, melt-H_2O content and/or melt composition and is frequently associated with magma recharge or mixing events. Pressure changes that exceed several Pascal are already capable of changing plagioclase composition (Nelson and Montana, 1992), which implies that the broad diffuse cores of the simple-zoned amphibole and oscillatory- zoned plagioclase phenocrysts then likely reflect crystallization under stable magma chamber conditions over longer periods of time (e.g. Foley et al., 2013). However, the presence of several successive zones with markedly increased An content in oscillatory-zoned plagioclase in the dacites and trachydacites (up to 15 mol %) likely reflects magma mixing and incoming batches of more mafic melts, in line with sieve-textured plagioclase phenocrysts that show An-rich overgrowths rims and embayments in quartz (when present). When hot and more mafic magma intrudes a colder, more differentiated reservoir, crystals near the mixing interface are re-melted and produce sieve textures and embayments (e.g. Donaldson and

Henderson, 1988; Tepley et al., 2001; Andrews et al., 2008), but ultimately causes more anorthitic plagioclase to grow onto e.g. the remaining plagioclase fragments (e.g. Troll et al., 2004). This interpretation is supported by sieved textures plagioclase cores in dacites that are overgrown by rims with elevated X_{An} (An_{44}) and Mg and Fe contents compared to their core (~An29).

The dacite/rhyolite domes in the southern part of the Sabzevar zone have high Sr/Y and La/Yb ratios but low Y (2.2-13 ppm) and Yb (0.5-1.5 ppm) contents, resembling typical adakites as defined by Defant and Drummond (1990). However, the Sr contents (209–377.5 ppm) for 13 of the 25 rhyolite samples are not as high as known from typical adakites (>400 ppm). Notably, oceanic slab-derived adakites are generally metaluminous, whereas the southern adakite rocks are metaluminous to peraluminous with low Mg# (Fig. 5d) and so differ from delamination derived-adakites. Compared to the adakite rocks derived from partial melting of subducted oceanic crust, most samples of southern domes are potassium-rich (K_2O up to 7 wt.%), with low A/NK ratios, and are thus akin to adakites of lower-crustal derivation (e.g. Wang et al., 2012) (Fig. 5 c), which may point to a thickened crust in the region. In contrast, a significant correlation between major and incompatible element ratios vs. SiO_2 between the southern high-silica samples and the northern intermediate rocks strengthen the hypothesis that the southern adakite samples are derived from the northern intermediate melts through continued crustal differentiation, e.g. assimilation and fractional crystallization. A co-evolution of northern intermediate magmas and southern felsic rocks is also supported by a decrease in the Dy/Yb ratios vs. SiO_2 (Fig. 6i) that implies that the northern intermediate magma experienced intense amphibole fractionation, which in turn will have helped to generate the high silica adakite-like magmatism in the southern part. In contrast, Rossetti et al. (2014) proposed wet amphibolite melting during subduction and then magma differentiation and high pressure amphibole fractionation of pristine slab melts for the genesis of Soltanabad adakite suites in the NE Sabzevar (east of our southern domes). However, a final model is not yet available, as the peraliminous character of the southern suite may indicate open system processes (crustal assimilation etc.) in addition to pure fractionation-controlled phenomena. This concept will have to be tested further using isotopic methods.

5.1 The magma plumbing system

Our amphibole, plagioclase-melt thermobarometery and clinopyroxene-melt barometry on the northern adakite rocks offer two distinct sets of pressures and temperatures, which indicates two separate levels of major crystallization and thus magma storage (~750 and ~350 MPa; Fig. 16). Amphibole formed at both levels, i.e. at pressures of 800 to 700 and 450 to 350 MPa, while the majority of plagioclase crystallized in a pressure range of 450 to 300 MPa. Little is known about structure of the crust beneath the Sabzevar ophiolitic zone, however, Motaghi et al. (2012) imaged the crust-mantle boundary beneath northeast Iran and Central Iran shows a strong variation of Moho depth from 35 km under Central Iran to 55km beneath the NE Iran. Following this model and that of Dehghani et al. (1983), we assume the Moho in the study region at ~ 45 km depth. Therefore, the Mg-hastingsites in the northern part (from Mg-andesite) record a mid- to lower crustal pressure (avg. ~700 MPa). In contrast, the other amphiboles (tschermakites and Mg-hornblendes) and plagioclases formed during a later stage of crystallization at a shallower level (~300 MPa). The absence of plagioclase in the Mg-andesite might indicate high contents of water in the magma source because the plagioclase is unstable under high water pressure (Wang et al., 2012). Dissolved water may have been ~5.5 wt.% in the magma source when hornblende and clinopyroxene formed as phenocrysts, likely too hydrous for plagioclase. However, the matrix is composed of plagioclase with anorthite percent up to 70%. The presence of high An plagioclases in the matrix is also consistent with the inference of a deeper magma source with high water content, based on experimental evidence and plagioclase stability (e.g., Sisson and Grove, 1993; Ramos et al., 2005). Therefore, the Mg-andesite dyke was likely derived from a mid-crustal water-rich mafic magma (Fig.17a, b). A significant mid- to lower crustal magma storage region is also supported by our clinopyroxene-melt barometry that shows pressures of ~550 to 700 MPa for clinopyroxenes from Mg-andesite.

The much more abundant medium-anorthite plagioclase phenocrysts (An_{53}-$_{30}$), in equilibrium with a trachyandesite and trachydacite compositions, in turn, formed in the upper crust (equal to 300 MPa pressure in average), in agreement with an upper storage region indicated by amphibole barometry (Figs 16, 17 b). Fractional crystallization of amphibole from parental magma in the mid- to lower crust thus likely led to evolved magmas (e.g. trachydacite) in the shallower levels and document a lively crustal evolution for the samples in question. Plagioclase phenocrysts that display sieve textured cores with an oscillatory zoned overgrowth provide evidence for hotter, mafic magma recharge and subsequent mixing processes, while mineral compositional trends suggest crystal fractionation also.

According to Chaussard and Amelung (2014), arcs with young and thin, ideally sedimentary crusts are more probable to develop shallow magma reservoirs (e.g. 1–3 km below the surface) than arcs with old and thick crusts. Although local effects like crustal heterogeneities such as fractures or sediment layering likely exert an additional control on the depths of magma storage, the shallow reservoir defined for our rock suites (300 MPa; ~6 Km) lies somewhat deeper than the top 3 km of the crust, which likely relates to the thicker crust in the region (ca 45 km) and the dominantly crystalline character of the ophiolitic basement.

7 Conclusions

The main conclusions of this study may be summarized as follows:

1. In the northern part of the Sabzevar ophiolitic zone (NE Iran), numerous post-ophiolite subvolcanic domes and dykes of Mg-andesite, trachyandesite, trachydacite, dacite compositions were intruded. In the southern part of the Sabzevar belt, post-ophiolite domes of dacite/rhyolite composition are exposed. Based on the whole rock geochemical features, the northern and southern post-ophiolite rocks have adakite-like affinities. The northern suite is calc-alkaline and metaluminous in nature, while the most southern samples are high-K calc-alkaline and metaluminous to peraluminous.

2. All selected amphiboles from northern samples are low-Al# (<0.21) and are found to be in equilibrium with high H_2O_{melt} values of 4.2 to 10.3 wt.% (± 0.5), as demonstrated in section 4.4, consistent with values assumed for ‘wet’ primitive calc-alkaline magma suites.

3. For the northern samples, the obtained T (992- 836°C ± 23.5) and P (820-145 MPa) ranges, based on amphibole thermobarometery, record two distinct magma storage regions or levels: Mg-hastingsite in andesite records mid- to lower crustal pressures (~700-900 MPa) that is in line with results from pyroxene-melt barometry. The other amphiboles show a shallow magma storage level equal to pressure of 300 MPa, which is, in turn, supported by plagioclase-melt barometry results.

4. The presence of amphibole and clinopyroxene without plagioclase phenocryst in Mg-andesite and the existence of amphibole clots in trachyandesites emphasises that fractionation of amphibole from high H_2O-magma was an important process in producing the diverse range of post-ophiolite rocks observed in the northern sector.

5. Our textural study documents that mafic magma recharge was driving the evolution of magmatic differentiation and disequilibrium features such as complex zoning and sieved textures in plagioclase, embayed quartz and zoned amphibole with high Mg# in outer margin and suggests that conditions changed for hotter and variably high fO_2 conditions due to magmatic replenishments.

6. Selected plagioclases from southern rhyolites (An_{8-15}) show pressures of crystallization between 273 to 110 MPa, implying a shallow upper crustal magma storage region for the southern sector also. According to the geochemical characteristics presented, we propose that the southern silica-rich peraluminous and high-K adakite-like rocks have been produced by differentiation of northern melts through amphibole fractionation, and possibly associated crustal assimilation (e.g. AFC) during upper crustal residence.

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Appendix:

P-T conditions of selected minerals were obtained using amphibole thermobarometry, plagioclase-liquid thermobarometry and clinopyroxene-liquid barometry. The equations which are used for the amphibole thermobarometry have been instructed by Ridolfi and Renzulli (2012), in detail. For plagioclase-liquid thermobarometry after Putirka (2008), T and P were obtained by equation (23) and equation (25a), respectively:

Equation (23):

$$“\frac{10^4}{\mathrm{T\ (K)}} = 6.12 + 0.257\ln\left(\frac{X_{\mathrm{An}}^{\mathrm{pl}}}{X_{\mathrm{CaO}}^{\mathrm{liq}}\,(X_{\mathrm{AlO_{1.5}}}^{\mathrm{liq}})^2 + (X_{\mathrm{SiO_2}}^{\mathrm{liq}})^2}\right) - 3.166\,(X_{\mathrm{CaO}}^{\mathrm{liq}}) - 3.137\left(\frac{X_{\mathrm{AlO_{1.5}}}^{\mathrm{liq}}}{X_{\mathrm{AlO_{1.5}}}^{\mathrm{liq}} + X_{\mathrm{SiO_2}}^{\mathrm{liq}}}\right) + 1.216(X_{\mathrm{Ab}}^{\mathrm{pl}})^2 -$$

$$2.475 \times 10^{-2}\ \left(\mathrm{P(kbar)} + 0.2166(\mathrm{H_2O^{liq}}\right)$$

Here, $X_{\mathrm{An}}^{\mathrm{pl}} = X_{\mathrm{CaO}}^{\mathrm{pl}}/(X_{\mathrm{CaO}}^{\mathrm{pl}} + X_{\mathrm{NaO_{0.5}}}^{\mathrm{pl}} + X_{\mathrm{KO_{0.5}}}^{\mathrm{pl}})$ and $X_{\mathrm{Ab}}^{\mathrm{pl}} = X_{\mathrm{NaO_{0.5}}}^{\mathrm{pl}}/(X_{\mathrm{CaO}}^{\mathrm{pl}} + X_{\mathrm{NaO_{0.5}}}^{\mathrm{pl}} + X_{\mathrm{KO_{0.5}}}^{\mathrm{pl}})$, where mineral components are calculated as cation fractions; All liquid components except H_2O are cation fractions. Calculate on an anhydrous basis without renormalization of weight percent values; H_2O is the wt% in the liquid phase.

Equation (25a) :

$$\mathrm{P\ (kbar)} = -42.2 + 4.94 \times 10^{-2}\ \mathrm{T(K)} + 1.16 \times 10^{-2}\mathrm{T(K)}\ln\left(\frac{X_{\mathrm{AlO_{1.5}}}^{\mathrm{liq}} X_{\mathrm{Ab}}^{\mathrm{pl}}\, X_{\mathrm{CaO}}^{\mathrm{liq}}}{X_{\mathrm{An}}^{\mathrm{pl}}\, X_{\mathrm{NaO_{0.5}}}^{\mathrm{liq}} X_{\mathrm{SiO_2}}^{\mathrm{liq}}}\right)$$

$$-382.3\,(X_{\mathrm{SiO_2}}^{\mathrm{liq}})^2 + 514.2\,(X_{\mathrm{SiO_2}}^{\mathrm{liq}})^3 - 19.6\ln(X_{\mathrm{Ab}}^{\mathrm{pl}}) - 139.8\,(X_{\mathrm{CaO}}^{\mathrm{liq}}) + 287.2\left(X_{\mathrm{NaO_{0.5}}}^{\mathrm{liq}}\right) + 163.9\,(X_{\mathrm{KO_{0.5}}}^{\mathrm{liq}})$$

For clinopyroxene-liquid barometry, the Equation (32c) which is presented by Putirka (2008) was used.

Equation (32c):

$$\mathrm{P\ (Kbar)} = -57.9 + 0.0475\mathrm{T\ (K)} - 40.6(X_{\mathrm{FeO}}^{\mathrm{liq}}) - 47.7\,(X_{\mathrm{CaTs}}^{\mathrm{cpx}}) + 0.676\,(\mathrm{H_2O^{liq}}) - 153\left(X_{\mathrm{CaO_{0.5}}}^{\mathrm{liq}} X_{\mathrm{SiO_2}}^{\mathrm{liq}}\right)$$

$$+ 6.89\left(\frac{X_{\mathrm{Al}}^{\mathrm{cpx}}}{X_{\mathrm{AlO_{0.5}}}^{\mathrm{liq}}}\right)$$

Here, the last term represents the ratio of the total number of Al in clinopyroxene when calculated on a 6 oxygen basis ($X_{Al}^{cpx} = X_{Al^{(IV)}}^{cpx} + X_{Al^{(VI)}}^{cpx}$), and the cation fraction of Al in liquid".

Figure Captions:

Figure.1 (a) Shaded relief image (after Shabanian et al., 2012) showing the location of Sabzevar Ophiolitic Zone (SOZ). Lower left inset shows the geodynamic setting of Arabia-Eurasia collision/subduction framework. Black arrows and associated numbers represent the present-day Arabia-Eurasia plate movement velocities after Reilinger et al. [2006] in mm/yr. (b) Simplified geological map of the Sabzevar area and the post-ophiolite subvolcanic rocks. The northern (dash square) and southern (dotted square) sector of the study area are indicated.

Figure.2 Field photographs. (a) and (b) northern grey dome of trachydacite exposed in ophiolite-related harzburgite. (b) Contact between post- ophiolite subvolcanic rock and host harzburgite. (c) Photograph of post-ophiolite dyke in the northern sector. (d) Presence of star-shape amphibole aggregates in the andesitic dyke. (e) and (f) photographs of southern post-ophiolite domes that outcrop in the ophiolite complex. Eocene volcaniclastic complex, Miocene sedimentary and Pliocene conglomerates are seen in (f).

Figure. 3 (a, b) Field photographs of host rock xenoliths. (a) Harzburgite belongs to ophiolite complex. (b) Eocene tuffaceous siltstone. (c) Rhyolitic fragments in the Pliocene conglomerate.

Figure. 4 Representative photomicrographs (XPL). (a) Photomicrograph of amphibole and clinopyroxene phenocrysts in a fine-grain matrix in andesite. (b) Flow texture and presence of acicular hornblende and plagioclase in andesite. (c) Euhedral to subhedral amphiboles with compositional zoning in trachy-andesite. (d) Photomicrograph showing contact between crystal clot containing plagioclase and amphibole and host trachyandesite rock. (e) Zoned plagioclase phenocryst in decite. (f) Embayment and rounded quartz, altered sanidine and plagioclase in rhyolite.

Figure. 5 Selected major element plots for the subvolcanic post-ophiolite rocks. (a) Total alkalis vs. silica diagram after Le Bas et al. (1986). The boundary between alkaline and subalkaline series is after Irvine and Baragar (1971), and the fields of low-silica adakites (LSA) and high-silica adakites (HSA) are after Martin et al., (2005). (b) K_2O vs. SiO_2 diagram (after Peccerillo and Taylor, 1976); (c) A/CNK vs. A/NK diagram (Maniar and Piccoli, 1989); (d) Mg# vs. SiO_2 diagram. The fields of subduction-related adakites and lower crust-derived adakites are from Guan et al. (2012).

Figure. 6 Harker diagrams of the major oxides (wt. %) and incompatible element ratios vs. SiO_2 for the studied rocks.

Figure. 7 Normalized REE and trace element patterns for the northern (a, b) and southern (c, d) subvolcanic rocks. Primitive-mantle and chondrite data are from Sun and McDonough (1989).

Figure. 8 (a) Plot of Sr/Y vs. Y for the studied rocks. Fields of adakite and arc normal rocks are from Petford and Atherton (1996) and Defant and Drummond (1990). (b) Plot of Sr vs. CaO+Na_2O (wt.%) showing the chemical differences between low silica adakites (dashed field) and high silica adakites (dotted field) (after Castillo, 2012).

Figure.9 Major element classification diagrams for amphiboles. (a) Amphiboles in the Mg-andesite samples. (b) Amphiboles in andesites with $^{V1}Al < Fe^{+3}$ fall in the Mg-hastingsite field.

(c) Amphiboles in the trachyandesite, trachydacite and dacite samples and (d) amphibole from crystal clots on the Mg/(Mg+Fe^{2+}) vs. Si classification diagrams (after Leake, 1978, Leake et al., 1997, 2004). (e) ^{V1}Al plotted against ^{IV}Al per formula unit and (f) (Na+K)^{A} vs. Si per formula unit for amphiboles from all samples. Procedure for assigning Al according to Leake (1978) and Leake et al. (2004). Solid black line in (e) indicates slope of 1.

Figure.10 (a) Back-scatter electron microprobe image of a representative amphibole phenocryst from a trachydacite sample that comprises distinct alternating dark and light zones. (b) and (c) Major element traverses in multiple- zoned amphibole phenocryst from (a). Arrow indicates position of rim to core traverse. Scale bars represent 100 μm.

Figure. 11 (a) Composition of all analyzed plagioclase data points (n= 212) in the feldspars composition triangle. Plagioclase composition in the samples of the northern sector vary between An_{20} and An_{70}. Plagioclase in rhyolites from the south part range from An_{20} to An_{8}. The BSE images of selected plagioclase phenocrysts show (b) oscillatory zoned plagioclase with abrupt increasing in An contents (up to 15 mol %) toward the rim and (c) sieve-textured plagioclase mantled by high An rim. (d) and (e) zoning profiles showing variations in X_{An} and MgO and FeO for the oscillatory zoned plagioclase and the sieve-textured plagioclase in (b) and (c), respectively. Yellow arrows indicate core to rim profile.

Figure. 12 Composition of clinopyroxene (n. 29). All clinopyroxene data plots in the Diopside field.

Figure. 13 (a) P-T diagram for calcic experimental amphiboles after the Ridolfi and Renzulli (2012) with low uncertainties (T ± 23.5°C, P ± 11.5%). Our amphiboles plot in the field of experimentally re-produced amphiboles (after Ridolfi and Renzulli 2012). (b) Log (fO_2)-T and (c) T-H_2O_{melt} diagrams for the studied amphiboles as obtained by amphibole thermobarometry after Ridolfi and Renzulli (2012). This method indicates typical uncertainties for log fO_2 errors (±0.2 log unit) and H_2O_{melt} (±0.8 wt%). The NNO and NNO +2 curves are from O'Neill and Pownceby (1993).

Figure.14 (a) Equilibrium test for plagioclase and four possible melts. The andesite samples appear to be in equilibrium with high anorthite plagioclase (An_{70-50}), while the trachyandesite/trachydacite and dacite are in equilibrium with (An_{50-30}) and (An_{30-20}), respectively. Rhyolite, in turn, is in equilibrium with low anorthite plagioclase (An_{20-10}). (b) Results of plagioclase-melt thermobarometry after (Putirka, 2008). SEE for the plagioclase-melt thermobarometer are ±36°C and ±247 MPa.

Figure.15 (a) Test for equilibrium using the Kd[FeMg] between clinopyroxene and nominal melt (basaltic andesite sample [no. N-1]). The result shows Kd[FeMg] values close to the ideal of 0.27 (Putirka 2008). (b) Test for equilibrium using predicted vs. observed clinopyroxene components of diopside + hedenbergite derived using the nominal equilibrium melt (basaltic andesite dyke). Pressure calculated for selected clinopyroxene using clinopyroxene-melt barometry after Putirka (2008).

Figure.16 A comparison of the results of amphibole, plagioclase-melt and clinopyroxene-melt barometries that indicate two distinct magma storage regions in the crust, one around a depth of ~6-9 km and the other at ~18-27 km depth.

Figure.17 (a) Tectonic framework illustrating northward subduction of Sabzevar oceanic crust (eastern branch of Neo-Tethys) beneath eastern Alborz zone. (b) Schematic illustration of the magma plumbing system for subvolcanic adakite-like rocks in northern part of Sabzevar belt based on the derived thermobarometric data.

Table 1. Major and trace element contents of representative post- ophiolite subvolcanic samples of NE Iran.

Samples	**N-1**	**N-2**	**N-3**	**N-4**	**N-5**	**N-6**	**N-7**	**N-8**	**N-9**	**N-10**	**N-11**	**N-12**	**N-13**	**N-14**	**N-15**
SiO_2	55.01	57.35	57.42	58.28	58.39	58.74	58.74	55.53	56.86	58.81	58.97	59.49	59.28	60.10	60.41
TiO_2	0.68	0.58	0.52	0.48	0.53	0.71	0.56	0.70	0.64	0.66	0.44	0.59	0.43	0.50	0.45
Al_2O_3	17.25	14.25	18.06	17.08	15.03	16.70	18.59	18.24	18.22	16.75	17.27	15.20	15.59	19.07	17.82
FeO	4.30	3.37	2.44	2.57	1.94	1.97	2.19	3.03	2.69	2.78	3.11	2.70	2.24	1.68	1.52
Fe_2O_3	2.87	3.37	4.53	2.57	1.94	1.97	4.06	4.55	4.03	1.85	2.07	2.70	1.49	3.12	2.28
MnO	0.14	0.13	0.13	0.15	0.18	0.21	0.12	0.11	0.11	0.24	0.16	0.13	0.17	0.04	0.06
MgO	4.23	6.03	3.91	3.25	2.83	3.82	2.61	4.98	4.38	2.76	1.86	3.74	4.56	2.06	4.62
CaO	5.21	5.38	7.06	3.96	3.51	1.50	6.44	4.31	4.40	5.25	5.63	3.08	4.59	5.21	4.12
Na_2O	5.27	5.26	3.91	5.42	4.80	5.47	4.06	6.05	6.68	4.15	4.13	6.97	4.38	3.98	4.93
K_2O	1.15	2.95	1.28	1.74	2.50	1.75	0.59	1.23	1.10	1.25	0.63	1.02	1.59	1.73	1.87
P_2O_5	0.13	0.11	0.19	0.11	0.06	0.07	0.16	0.11	0.15	0.18	0.10	0.10	0.07	0.18	0.15
L.O.I.	1.8	0.5	2.10	2.1	4.2	3.5	3.30	2.70	2.10	2.6	2.8	2.1	2.7	3.70	3.10
Sum	99.87	99.85	101.5	99.89	99.93	101.43	99.89	101.63	101.33	99.89	99.93	99.90	99.93	101.35	101.91
Ba	122	96	357.7	143	288	470	130.1	135.7	149.7	238	153	102	270	206.0	381.0
Co	44.3	54.3	21.90	35.6	28.6	30.5	13.10	24.80	21.00	46.7	27.4	18.9	44.2	8.70	15.60
Cs	0.1	0.1	0.60	0.5	1.4	1.4	0.40	1.00	0.10	1.6	0.2	0.2	0.6	0.50	0.60
Ga	16.82	12.69	17.40	16.8	14.02	16.41	17.70	16.40	17.50	16.72	16.85	14.67	15.82	16.30	15.80
Hf	1.8	1.62	2.00	0.94	1.22	1.49	2.20	2.10	2.20	2.89	1.18	2.41	1.11	2.70	2.90
Nb	5	5	2.00	5	5	8	2.70	1.70	1.90	8	5	5	5	4.10	4.80
Rb	4	12.4	19.80	12.7	39.1	18.5	9.20	22.60	10.20	17.6	3.1	5.7	7.7	22.40	31.80
Sr	390	64	780.6	350	403	465	366.1	463.9	208.6	476	407	138	783	470.0	731.7
Ta	0.2	0.3	0.10	0.3	0.4	0.6	0.20	0.10	0.20	0.5	0.3	0.2	0.5	0.30	0.40
Th	0.5	0.6	2.50	0.3	1.6	1.2	1.30	1.10	0.90	2.2	0.5	0.8	1.2	1.50	3.00
U	0.2	0.1	0.80	0.1	0.5	0.6	0.40	0.30	0.40	0.8	0.1	0.3	0.3	0.50	1.20
V	176	167	208.00	111	76	80	134.00	207.00	170.00	117	87	119	82	70.00	76.00
Zr	81	66	69.20	87	109	124	72.60	64.60	66.60	132	87	93	110	110.60	103.30
Y	17	14	11.70	14	10	11	14.60	17.30	17.10	14	14	19	8	14.50	8.70
La	3.3	4.7	10.90	3.7	6	5.8	7.50	5.80	5.60	7.8	5	4.3	7.3	8.80	11.90
Ce	9.13	11.31	23.40	9.97	14.03	14.18	17.20	13.10	13.00	18.36	11.46	11.1	16	21.40	22.20
Pr	1.4	1.7	2.94	1.4	1.9	1.9	2.31	1.92	1.96	2.3	1.6	1.9	2	2.70	2.66
Nd	6.2	7.5	12.70	6	7.2	7.4	10.70	9.60	9.30	8.3	7	7.1	7	10.90	10.60
Sm	1.9	2.3	2.60	1.7	1.7	1.7	2.30	2.70	2.60	1.9	1.7	2.2	1.5	2.50	2.10
Eu	0.8	0.8	0.79	0.7	0.5	0.6	0.72	0.77	0.80	0.6	0.6	0.8	0.4	0.79	0.59
Gd	2.7	2.8	2.16	1.7	1.7	1.6	2.81	3.12	3.02	2.1	2	2.4	1.1	2.67	1.97
Tb	0.4	0.4	0.38	0.3	0.3	0.2	0.45	0.55	0.52	0.4	0.3	0.5	0.3	0.43	0.30
Dy	2.3	2.8	2.06	1.5	1.5	1.7	2.62	3.21	2.80	2.5	1.9	2.6	1.1	2.54	1.63
Ho	0.6	0.5	0.37	0.4	0.3	0.3	0.50	0.61	0.57	0.5	0.3	0.7	0.2	0.43	0.29
Er	1.5	1.7	1.14	1	0.9	0.7	1.50	1.76	1.84	1.2	0.9	1.9	0.6	1.35	0.88
Yb	1.3	1.6	1.10	0.9	0.8	0.7	1.54	1.67	1.73	1.3	0.9	2	0.6	1.35	0.84
Lu	0.2	0.3	0.18	0.1	0.1	0.1	0.24	0.29	0.28	0.2	0.1	0.3	0.1	0.23	0.14
Sr/Y	22.94	4.57	66.71	25.00	40.30	42.27	25.07	26.81	12.19	34.00	29.07	7.26	97.88	32.41	84.10
La/Yb	2.54	2.94	9.91	4.11	7.50	8.29	4.87	3.47	3.24	6.00	5.56	2.15	12.17	6.52	14.17
Zr/Sm	42.63	28.70	4.05	51.18	64.12	72.94	5.70	5.82	6.43	69.47	51.18	42.27	73.33	9.48	8.69
Mg#	50	59.3	61.6	56	67.1	66	54.4	62.2	62	50	37.4	58.1	67.1	55.1	75.2
Eu/Eu*	1.1	1	1	1.2	0.9	1	0.9	0.8	0.9	0.9	1	1.1	1	0.9	0.9
A/CNK	0.8	0.6	0.8	0.9	0.8	1.2	0.9	0.9	0.9	1	0.9	0.8	0.8	1.1	1

Table 1 (continued)

Samples	**N-16**	**N-17**	**N-18**	**N-19**	**N-20**	**N-21**	**N-22**	**S-23**	**S-24**	**S-25**	**S-26**	**S-27**	**S-28**	**S-29**	**S-30**
SiO_2	62.56	60.1	62.31	64.1	62.78	68.61	63.60	63.71	69.72	70.23	70.28	70.40	70.72	70.78	70.95
TiO_2	0.48	0.48	0.42	0.32	0.41	0.15	0.47	0.31	0.02	0.04	0.10	0.04	0.07	0.07	0.08
Al_2O_3	17.74	17.13	15.51	17.04	18.31	18.82	17.29	16.78	20.64	20.24	18.73	19.81	18.67	16.60	18.2
FeO	1.78	2.43	2.33	1.53	1.63	0.81	1.80	1.88	0.25	0.30	0.51	0.28	0.41	0.36	0.44
Fe_2O_3	2.68	1.62	1.55	1.53	1.63	0.81	2.70	1.25	0.25	0.30	0.51	0.28	0.41	0.36	0.44
MnO	0.09	0.22	0.12	0.20	0.05	0.08	0.08	0.15	0.06	0.04	0.04	0.04	0.03	0.06	0.04
MgO	3.06	2.17	2.96	1.17	2.45	0.75	2.92	2.10	0.10	0.18	0.25	0.16	0.24	0.22	0.23
CaO	5.41	4.38	4.26	3.31	3.62	0.87	5.25	4.59	1.25	0.46	1.97	0.42	0.81	2.68	1.81
Na_2O	4.39	4.58	4.19	4.69	5.27	6.51	4.51	4.35	4.57	5.38	5.33	5.37	5.73	4.85	5.26
K_2O	0.72	1.54	1.35	2.13	2.05	2.24	0.77	1.11	3.57	3.62	2.90	3.48	3.35	1.92	2.67
P_2O_5	0.11	0.06	0.08	0.09	0.22	0.13	0.13	0.06	0.03	0.05	0.04	0.03	0.03	0.06	0.06
L.O.I.	2.40	2.6	2.4	1.9	3.10	1.60	1.90	1.8	1.10	0.70	0.80	1.10	1.00	3.50	1.20
Total	101.41	99.94	99.98	99.95	101.43	101.32	101.4	99.93	101.50	101.53	101.41	101.41	101.44	101.43	101.34
Ba	130.3	272	181	300	445.6	463.9	132.5	323	518.0	545.1	481.0	559.6	539.2	318.5	413.4
Co	13.50	22.2	49.5	26.6	9.30	1.20	13.60	31.8	2.00	2.00	2.00	1.00	2.00	1.00	1.00
Cs	0.20	0.4	0.4	1	0.90	0.80	0.20	0.7	0.5	0.50	1.00	0.5	1.10	1.20	1.30
Ga	15.90	15.97	16.28	15.36	16.80	17.70	15.80	15.71	2.60	1.80	0.80	1.80	17.40	17.20	17.80
Hf	2.40	2.69	1.95	0.58	3.20	2.70	2.80	0.9	18.70	16.10	16.40	16.00	1.90	2.10	2.40
Nb	2.40	9	8	15	5.80	7.70	2.60	5	2.10	1.90	2.00	1.90	5.20	5.40	5.00
Rb	10.10	9.3	11.6	42.4	38.70	43.10	10.5	10.9	6.80	3.70	5.00	4.10	65.50	35.70	52.70
Sr	507.1	550	283	358	848.5	673.5	400	670	70.10	73.60	58.00	68.30	811.00	512.2	810.6
Ta	0.10	0.6	0.4	0.9	0.40	0.60	0.10	0.5	337.80	344.50	894.50	335.10	0.40	0.30	0.30
Th	0.80	3.6	1.4	4.7	3.60	3.80	0.70	1.1	0.60	0.30	0.30	0.30	0.50	0.70	0.70
U	0.40	0.9	0.5	1	1.20	1.20	0.40	0.2	1.60	1.70	1.50	1.50	1.10	0.70	1.10
V	86.00	64	88	31	58.00	19.00	90.00	68	1.80	1.00	1.10	1.00	6.00	7.00	7.00
Zr	85.70	160	92	186	124.30	90.10	88.00	71	5	5	9.00	5	46.40	50.0	53.8
Y	16.80	12	10	15	9.20	9.20	17.00	5	38.40	40.50	61.40	38.80	2.90	2.70	3.20
La	6.10	12.6	4.9	15.9	13.00	15.40	6.90	6.1	6.30	5.70	3.40	5.50	2.70	2.50	3.40
Ce	15.80	26.61	11.37	35.1	26.40	31.50	16.30	12.27	3.80	5.50	5.10	5.90	4.70	4.60	6.80
Pr	2.25	3	1.7	4	3.04	3.60	2.34	1.4	7.50	10.10	9.20	10.20	0.63	0.59	0.82
Nd	10.30	10.9	6.2	13.6	11.10	12.90	10.10	5.9	0.93	1.21	1.14	1.29	2.70	2.20	3.20
Sm	2.50	1.8	1.7	2.4	2.10	1.90	2.50	1.5	3.50	4.30	4.20	5.10	0.50	0.40	0.60
Eu	0.72	0.7	0.6	0.6	0.61	0.55	0.69	0.4	0.80	1.00	0.80	1.00	0.14	0.16	0.18
Gd	2.80	1.9	1.7	1.7	1.59	1.59	2.83	0.8	0.26	0.25	0.24	0.29	0.39	0.38	0.50
Tb	0.49	0.3	0.3	0.4	0.27	0.26	0.48	0.1	1.02	0.96	0.61	0.97	0.08	0.07	0.09
Dy	2.88	1.4	1.6	1.7	1.46	1.39	2.74	0.7	0.18	0.17	0.10	0.16	0.39	0.39	0.45
Ho	0.54	0.4	0.4	0.4	0.29	0.26	0.59	0.1	0.96	0.77	0.54	0.91	0.08	0.07	0.09
Er	1.82	1	1	0.9	0.86	0.78	1.70	0.4	0.18	0.16	0.10	0.16	0.21	0.22	0.27
Yb	1.67	1.1	0.8	0.7	0.77	0.78	1.73	0.4	0.52	0.43	0.25	0.41	0.17	0.18	0.22
Lu	0.29	0.2	0.1	0.1	0.14	0.13	0.30	0.1	0.47	0.38	0.26	0.42	0.03	0.03	0.03
Sr/Y	30.1	45.83	28.30	23.87	92.22	73.20	23.53	134.00	15.96	26.58	35.38	24.29	279.6	189.7	253.3
La/Yb	3.65	11.45	6.13	22.71	16.88	19.74	3.99	15.25	3.99	53.61	263.0	60.92	15.88	13.89	15.45
Zr/Sm	7.27	88.89	54.12	77.50	15.13	161.55	7.22	47.33	549.69	1086	130	1230	381	102	153.79
Mg#	63.2	47.2	56	43.4	60	48.1	62	52.8	28.6	37.5	33	36.4	37	37	34.3
Eu/Eu*	0.8	1.1	1.1	0.9	1	1	0.8	1.1	0.9	.0.8	1.1	0.9	1	1.2	1
A/CNK	1	1	1	1	1	1.1	0.9	1.2	2.2	1.2	1.1	1.2	1.1	1.2	1

Table 1 (continued)

Samples	**S-31**	**S-32**	**S-33**	**S-34**	**S-35**	**S-36**	**S-37**	**S-38**	**S-39**	**S-40**	**S-41**	**S-42**	**S-43**	**S-44**	**S-45**
SiO_2	71.27	70.7	71.3	71.1	71.59	72.3	72.59	73.2	72.09	72.30	72.32	72.46	72.54	72.64	72.92
TiO_2	0.07	0.08	0.08	0.10	0.02	0.03	0.04	0.02	17.64	17.53	17.62	16.88	16.94	17.23	17.52
Al_2O_3	17.14	15.4	16.6	16.3	15.24	15.5	14.90	16.4	0.32	0.34	0.38	0.34	0.28	0.37	0.35
FeO	0.42	0.40	0.66	0.41	0.33	0.36	0.34	0.39	0.32	0.34	0.38	0.34	0.28	0.37	0.35
Fe_2O_3	0.24	0.49	0.66	0.50	0.40	0.44	0.42	0.39	0.12	0.18	0.24	0.13	0.14	0.18	0.16
MnO	0.05	0.05	0.15	0.05	0.09	0.09	0.04	0.08	1.52	0.40	0.26	0.51	0.80	0.60	1.34
MgO	0.43	0.25	0.53	0.26	0.20	0.25	0.21	0.13	5.12	6.50	7.07	5.57	5.27	6.22	5.58
CaO	0.35	1.75	0.97	1.85	0.37	1.26	0.42	1.19	3.48	3.10	2.65	4.21	3.55	3.17	2.36
Na_2O	6.32	5.13	5.27	5.50	5.80	5.18	5.43	5.10	0.03	0.05	0.05	0.02	0.02	0.05	0.01
K_2O	3.46	2.62	3.50	2.76	2.91	2.56	3.47	3.75	0.06	0.04	0.06	0.03	0.01	0.06	0.01
P_2O_5	0.06	0.05	0.17	0.05	0.08	0.07	0.05	0.05	0.06	0.03	0.03	0.06	0.09	0.03	0.07
L.O.I.	0.80	1.4	1.40	0.4	1.4	0.9	1	0.80	0.90	0.70	0.60	1.00	1.50	0.60	0.80
Total	100.6	99.9	101.	99.9	99.96	99.9	99.98	101.	101.6	101.5	101.6	101.52	101.4	101.5	101.4
Ba	424.3	408	558.	416	541	471	479	493.	507.5	466.5	445.5	510.1	502.8	487.8	310.8
Cs	0.50	1.3	1.30	0.4	0.9	0.6	2.3	3.20	2.50	0.70	0.50	1.30	2.00	1.20	1.20
Ga	16.50	17.9	17.4	17.2	17.66	17.6	16.22	16.9	18.20	18.40	16.90	18.10	16.80	17.20	18.30
Hf	2.00	0.96	3.20	0.66	0.96	1.09	1.18	2.10	2.00	1.90	1.40	2.10	2.10	1.90	1.90
Nb	5.20	5	9.40	5	12	7	5	9.30	7.20	6.40	6.70	9.30	7.10	6.30	10.60
Rb	65.30	60	70.9	49.3	60.4	52.7	68.2	83.1	72.70	67.80	52.40	92.10	71.30	65.90	46.00
Sr	208.2	774	338.	790	350	221	362	365.	475.5	537.7	377.5	459.9	246.7	644.3	283.2
Ta	0.40	0.6	0.80	0.6	1	0.7	0.6	0.70	0.60	0.50	0.40	0.80	0.70	0.50	1.00
Th	0.80	0.9	4.80	1	3.1	2.7	0.8	1.70	1.20	1.00	0.10	1.00	2.10	1.20	0.40
U	1.30	0.8	2.60	0.7	1.5	1.3	1.3	1.90	1.40	1.70	1.10	2.40	2.00	1.50	1.90
V	8.00	6	10.0	7	1	1	2	<5	<5	5.00	<5	<5	<5	<5	<5
Zr	49.0	59	74.0	67	52	45	41	39.5	41.1	43.9	31.60	39.4	41.9	43.70	24.80
Y	2.70	3	13.1	3	7	7	5	3.60	5.60	4.70	2.20	3.40	6.40	4.30	4.80
La	2.40	3.3	15.6	4.3	9.6	8.9	1.8	5.10	5.10	4.40	1.00	4.60	4.10	3.40	3.00
Ce	4.50	7.05	30.8	9.12	20.09	17.5	4.07	9.90	10.0	7.10	1.40	9.00	7.90	6.90	4.70
Pr	0.53	0.9	3.72	1	2.2	2	0.5	1.19	1.19	0.91	0.17	1.08	0.98	0.84	0.61
Nd	2.10	3.3	13.4	3.7	8.4	7.2	1.9	5.00	4.70	3.60	0.70	4.10	3.30	3.60	2.50
Sm	0.40	0.5	2.50	0.6	1.4	1.3	0.3	1.00	0.90	0.80	0.20	0.80	1.00	0.70	1.00
Eu	0.16	0.2	0.42	0.2	0.4	0.3	0.2	0.24	0.24	0.27	0.13	0.24	0.23	0.26	0.30
Gd	0.38	0.5	2.22	0.5	1.2	1.5	0.7	0.94	0.87	0.78	0.28	0.83	1.04	0.64	1.23
Tb	0.08	0.1	0.37	<0.1	0.2	0.2	0.1	0.14	0.17	0.14	0.06	0.14	0.18	0.14	0.19
Dy	0.37	0.5	2.15	0.3	1	1.1	0.7	0.72	0.85	0.71	0.36	0.63	0.96	0.68	0.89
Ho	0.08	0.1	0.37	0.1	0.2	0.2	0.2	0.09	0.14	0.12	0.05	0.07	0.16	0.12	0.11
Er	0.20	0.2	1.07	0.1	0.4	0.7	0.4	0.17	0.41	0.33	0.13	0.14	0.45	0.30	0.19
Yb	0.19	0.1	1.06	0.1	0.4	0.6	0.5	0.11	0.40	0.33	0.05	0.10	0.46	0.32	0.08
Lu	0.03	0.1	0.17	0.1	0.1	0.1	0.1	0.01	0.06	0.05	0.01	0.01	0.07	0.05	0.01
Sr/Y	77.11	258	25.8	263.	50	31.5	72.4	101.	84.91	114.4	171.5	135.2	38.54	149.8	59
La/Yb	12.63	33	14.7	43	24	14.8	3.6	46.3	12.75	13.33	20.00	46.00	8.91	10.63	37.50
Zr/Sm	723.6	118	147.	111.	37.1	34.6	136.7	244.	259.1	844.2	1137	673.1	498.2	510.1	263.3
Mg#	50.6	38.2	44.5	38.8	37.5	30.8	37.7	25	27.3	34.6	38.7	27.7	33.3	32.7	31.4
Eu/Eu*	1.1	1.2	1.2	1.1	1	0.8	1.2	0.8	0.9	1.1	1.6	0.9	0.8	1.2	0.9
A/CNK	1.1	1	1	1	1	1.2	1.1	1.6	1.1	1.2	1	1.2	1.2	1.1	1.1

Table 1 (continued)

Samples	**S-46**	**S-47**	**S-48**	**S-49**	**S-50**	**S-51**	**S-52**	**S-53**	**S-54**	**S-55**	**S-56**	**S-57**
SiO_2	73.14	73.62	67.54	70.12	65.53	65.78	67.15	69.97	66.83	64.91	67.89	66.39
TiO_2	17.55	0.03	0.38	0.26	0.44	0.44	0.38	0.29	0.38	0.48	0.34	0.38
Al_2O_3	0.36	15.19	16.59	16.73	17.37	16.95	16.61	15.84	16.63	16.92	16.26	17.36
FeO	0.36	0.25	1.30	0.89	1.47	1.47	1.29	0.95	1.32	1.71	1.15	1.29
Fe_2O_3	0.16	0.31	1.91	1.34	2.21	2.20	1.94	1.43	1.98	2.56	1.72	1.94
MnO	1.63	0.05	0.06	0.03	0.05	0.05	0.06	0.04	0.05	0.04	0.06	0.03
MgO	5.17	0.15	1.86	0.68	2.13	2.47	2.20	1.76	1.98	2.87	2.50	1.66
CaO	2.32	0.92	4.20	3.48	4.83	4.80	4.37	4.08	4.16	4.79	4.26	5.08
Na_2O	0.01	5.35	4.13	3.45	4.76	4.73	4.29	3.98	4.09	4.72	4.16	4.96
K_2O	0.01	3.33	1.45	1.58	1.30	1.36	1.44	1.34	1.82	1.33	1.60	1.34
P_2O_5	0.06	0.04	0.15	0.14	0.21	0.20	0.15	0.07	0.13	0.19	0.14	0.20
L.O.I.	0.70	0.3	1.51	0.75	1.41	1.45	1.79	2.33	1.64	1.48	2.21	2.4
Total	101.45	99.97	99.96	99.94	99.95	99.96	99.97	99.96	99.95	99.97	99.97	99.97
Ba	304	508	328	386	318	295	332	383	442	277	399	338
Cs	1.70	2.2	0.60	0.39	0.80	0.79	0.60	0.90	1.18	0.60	0.90	0.79
Ga	19.80	17.5	14.9	9.8	15.9	13.9	11.9	12.9	12.8	18.9	13.0	18.9
Hf	1.70	1.01	2.0	2.0	2.0	2.0	2.0	2.0	2.0	2.0	2.0	2.0
Nb	10.60	9	9	2.9	5.9	6.9	5	3	4	7.9	6	8
Rb	47.50	74.5	20.2	18.9	18.9	17.3	17.3	24.2	22.6	18.6	26.9	24.3
Sr	314.10	523	668.7	677.1	697	770.9	633.2	486.5	606.5	657.7	549	626.6
Ta	1.00	0.9	0.5952	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5	0.5
Th	0.30	1.7	2.4	2.2	1.8	2.5	2.8	1.8	3	3	3	2.2
U	1.70	1.7	0.84	0.76	0.62	0.83	0.81	0.80	0.99	0.83	0.96	0.82
V	<5	1	61.37	14.32	90.44	86.85	69.24	76.51	65.29	92.6	66.5	62.08
Zr	24.7	42	80.	61.8	83.5	66.5	61.6	66.7	55.9	93.5	81	109.2
Y	4.90	5	5	6	7	7	6	6	6	8	6	10
La	2.90	4.8	31.4	31.6	27.0	31.5	28.8	23.5	31.0	30.6	29.0	32.2
Ce	5.30	9.53	23.8	26.5	20.9	26.8	22.9	14.9	25.5	24.9	22.0	26.8
Pr	0.63	1.2	2.7	3.1	2.6	3.2	2.6	1.7	3.0	3.0	2.4	3.2
Nd	2.50	4.6	9.5	11.8	9.9	11.9	9.7	6.7	10.8	10.9	8.8	11.9
Sm	1.00	0.9	1.8	1.8	1.7	2.0	1.5	1.2	1.9	1.7	1.5	2.3
Eu	0.32	0.2	0.6	0.6	0.7	0.7	0.6	0.5	0.6	0.7	0.5	0.7
Gd	1.20	0.7	2.2	2.2	2.1	2.2	2.0	1.5	2.2	2.0	1.9	2.3
Tb	0.21	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Dy	1.03	0.8	0.8	1.1	0.7	1.0	0.5	0.8	1.0	1.2	0.8	0.7
Ho	0.11	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Er	0.21	0.3	0.6	0.7	0.5	0.6	0.5	0.5	0.7	0.6	0.6	0.6
Yb	0.11	0.3	0.7	0.9	1.0	1.0	1.0	1.0	1.0	1.0	1.2	1.6
Lu	0.01	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1	0.1
Sr/Y	64.10	104.6	133.7	112.9	99.5	110.1	105.5	81.1	101.1	82.2	91.5	62.7
La/Yb	26.36	16	14.7	25.8	27.0	28.8	29.0	30.6	31.4	32.2	35.1	45.1
Zr/Sm	212.31	46.67	45.0	35.0	49.4	33.5	41.3	55.8	30.0	55.3	54.0	47.8
Mg#	30.8	37.3	58.9	43.0	59.1	62.7	63.0	65.0	60.0	62.7	68.5	56.2
Eu/Eu*	0.9	0.8	1.0	1.0	1.1	1.1	1.0	1.1	0.9	1.2	0.9	0.9
A/CNK	1.2	1.2	0.9	1	1.2	1.1	1	0.9	0.9	0.9	1	1

Table 2. Representative compositions of amphiboles from post-ophiolite rocks in Sabzevar zone.

Mineral	**Unzoned-Amphibole**		**Simple-zoned Amphibole**		**Simple-zoned Amphibole**		**Simple-zoned Amphibole**		**Oscillatory-zoned Amphibole**			**Amphibole in crystal clots**		
Host rock	**N-1** (andesite)		**N-11** (Trachyandesite)		**N-12** (Trachydacite)		**N-12**(Trachydasite)		**N-4** (Trachydacite)			**N-6** (Trachyandesite)		
position	**Core**	**Rim**	**Core**	**Rim**	**Core**	**Rim**	**Core**	**Rim**	**Core**	**Bright zone**	**Rim**	**Rim**	**Core**	**Rim**
SiO_2	41.03	40.71	44.19	46.18	42.6	45.31	45.66	47.68	46.09	47.73	44.98	43.19	44.63	43.79
TiO_2	3.55	3.12	1.56	1.42	1.24	1.62	1.33	2.40	1.66	1.03	1.45	1.91	2.19	2.28
Al_2O_3	12.77	12.58	11.81	11.1	13.43	11.16	12.04	8.65	11.52	8.79	11.11	11.89	10.96	11.86
FeO	10.11	9.89	13.44	11.06	15	11.87	9.26	11.55	9.15	12.25	13.06	12.99	12.61	12.69
MnO	0.06	0	0.26	0.14	0.24	0	0.09	0.23	0.23	0.32	0.25	0.06	0.11	0.00
MgO	14.53	14.95	13.37	16.02	11.36	15.17	16.54	15.06	16.65	14.77	14.02	14.15	14.31	13.79
CaO	11.99	12.05	11.69	10.75	10.58	10.71	11.07	10.28	11.04	11.19	11.55	11.16	11.12	11.06
Na_2O	2.45	2.61	2.11	2.31	2.41	2.33	2.28	2.22	2.31	1.68	2.3	2.09	2.09	2.22
K_2O	0.96	0.99	0.16	0.16	0.30	0.12	0.36	0.34	0.2	0.19	0.18	0.28	0.29	0.37
Cr_2O_3	0.03	0.12	0.07	0.06	0.01	0.01	0.00	0.05	0.07	0.04	0.02	0.00	0.01	0.01
Total	97.47	97.02	98.67	99.21	97.18	98.30	98.63	98.57	98.95	98.00	98.93	97.73	98.33	98.07
Amphibole Formula on the basis of 13 cations (Leake et al., 1997)														
Si	6.00	5.98	6.41	6.41	6.22	6.39	6.36	6.74	6.40	6.79	6.41	6.19	6.36	6.28
Al^{iv}	2.00	2.02	1.59	1.59	1.78	1.61	1.64	1.26	1.60	1.21	1.59	1.81	1.64	1.72
Al^{vi}	0.20	0.15	0.23	0.23	0.53	0.24	0.34	0.18	0.28	0.26	0.28	0.20	0.21	0.28
Ti	0.39	0.34	0.15	0.15	0.14	0.17	0.14	0.25	0.17	0.11	0.16	0.21	0.23	0.25
Fe^{+3}	0.39	0.44	1.21	1.21	0.93	1.14	1.04	0.79	1.02	0.82	0.80	1.13	0.93	0.86
Mg	3.17	3.27	3.32	3.32	2.47	3.19	3.43	3.17	3.45	3.13	2.98	0.43	0.57	0.66
Fe^{+2}	0.84	0.77	0.08	0.08	0.90	0.26	0.04	0.57	0.04	0.63	0.75	0.01	0.01	0.00
Mn	0.01	0.00	0.02	0.02	0.03	0.00	0.01	0.03	0.03	0.04	0.03	3.03	3.04	2.95
Ca	1.88	1.90	1.79	1.60	1.65	1.62	1.65	1.56	1.64	1.70	1.76	1.71	1.70	1.70
Na	0.57	0.64	0.38	0.22	0.34	0.25	0.27	0.16	0.26	0.17	0.40	0.58	0.58	0.62
K	0.18	0.19	0.03	0.03	0.06	0.02	0.06	0.06	0.04	0.03	0.03	0.05	0.05	0.07

Table 3. Representative compositions of plagioclases from post-ophiolite rocks in Sabzevar zone.

Mineral	**Plagioclase**		**Normal- zoned Plagioclse**		**Oscillatory-zoned Plagioclase**			**Sieved texture Plagioclase**		**Plagioclase**	
Host rock	**N-10 (Andesite)**		**N-11 (Trachyandesite)**		**N-17 (dacite)**			**N-18 (dacite)**		**S-37 (Rhyolite)**	
Position	**Core**	**Rim**	**Core**	**Rim**	**Core**	**inner zone**	**Rim**	**Core**	**Rim**	**Core**	**Rim**
SiO_2	51.37	53.32	58.99	61.88	63.58	54.97	60.38	67.11	58.68	67.66	68.50
TiO_2	0.08	0.04	0.07	0.01	0	0.054	0	0.02	0.02	0	0.00
Al_2O_3	30.63	29.62	26.90	24.80	23.72	28.27	25.64	22.22	27.07	21.58	20.54
FeO	0.58	0.55	0.18	0.14	0.08	0.22	0.17	0.12	0.41	0	0.05
MnO	0.00	0.00	0.00	0.03	0	0.03	0.03	0	0	0.1157	0.18
MgO	0.08	0.06	0.01	0.00	0.02	0	0	0.04	0.02	0.0089	0.00
CaO	14.04	13.00	9.15	7.03	5.61	11.32	8.09	4.09	9.54	2.63	2.32
Na_2O	3.25	3.88	4.57	5.99	7.34	5.1	5.64	6.59	5.69	7.95	7.45
K_2O	0.08	0.15	0.08	0.12	0.36	0.14	0.26	0.46	0.23	0.4767	0.55
Cr_2O_3	0.03	0.00	0.00	0.01	0	0	0	0	0	0.0074	0.00
Total	100.14	100.61	99.96	100.01	100.72	100.11	100.22	100.67	101.66	100.4287	99.59
Feldspar Components											
An	0.70	0.65	0.52	0.39	0.29	0.55	0.43	0.25	0.47	0.15	0.14
Ab	0.29	0.34	0.47	0.60	0.69	0.45	0.55	0.72	0.51	0.82	0.82
Or	0.00	0.01	0.01	0.01	0.02	0.01	0.02	0.03	0.01	0.03	0.04

Table 4. Representative compositions of clinopyroxenes from post-ophiolite rocks in Sabzevar zone.

Mineral	**Cpx**								
Host rock	**N-1 (andesite)**		**N-1 (andesite)**		**N-1 (andesite)**			**N-1 (andesite)**	
Position	**Core**	**Rim**	**Core**	**Rim**	**Rim**	**Core**	**Rim**	**Core**	**Rim**
SiO_2	48.82	49.01	53.27	49.02	49.72	48.94	52.7	53.58	53.54
TiO_2	1.33	1.28	0.53	0.84	1.19	1.44	0.32	0.33	0.36
Al_2O_3	5.16	4.40	2.29	2.81	4.41	4.91	2	1.98	2.31
FeO	7.08	7.77	5.46	6.66	7.04	7.41	6.06	5.92	6
MnO	0.06	0.22	0.09	0.22	0.15	0	0.21	0.21	0.3
MgO	13.68	14.49	16.49	15.49	14.64	13.67	16.52	16.69	16.45
CaO	22.89	21.79	22.26	22.07	21.91	22.91	21.5	21.38	21.16
Na_2O	0.34	0.39	0.40	0.42	0.49	0.42	0.43	0.37	0.40
K_2O	0.00	0.02	0.30	0.00	0.02	0.00	0.01	0.00	0.02
Cr_2O_3	0.01	0.00	0.30	0.02	0.05	0.00	0.25	0.29	0.26
Total	99.37	99.35	101.12	97.55	99.96	99.71	100.00	100.76	100.79
Si	1.77	1.94	1.94	1.78	1.81	1.78	1.91	1.95	1.95
Ti	0.04	0.01	0.01	0.02	0.03	0.04	0.01	0.01	0.01
Al	0.17	0.07	0.07	0.09	0.14	0.16	0.06	0.08	0.07
Al^{iv}	0.23	0.06	0.06	0.22	0.19	0.22	0.08	0.05	0.05
Al^{vi}	0.01	0.01	0.01	0.02	0.05	0.06	0.02	0.03	0.02
Fe^{2+}	0.22	0.17	0.17	0.21	0.22	0.23	0.18	0.18	0.18
Mn	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.01	0.01
Mg	0.76	0.89	0.89	0.88	0.81	0.76	0.90	0.90	0.89
Ca	0.92	0.87	0.87	0.90	0.88	0.92	0.85	0.83	0.82
Na	0.02	0.03	0.03	0.03	0.04	0.03	0.03	0.02	0.03
Mg#	0.77	0.84	0.84	0.81	0.79	0.78	0.83	0.83	0.83