

**Post-collisional ultrapotassic volcanic rocks and ultramafic xenoliths in the Eslamieh Peninsula,
NW Iran: Petrological and geochemical constraints on mantle source and metasomatism**

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Abstract

The Eslamieh Peninsula in NW Iran exposes Miocene potassic (or shoshonitic) to ultrapotassic volcanic rocks that are often associated with a variety of mantle (pyroxenite and glimmerite) xenoliths. Based on geochemical features, the rocks investigated in this study can be subdivided into two main groups: 1) a MgO-rich (8.0-13.9 wt. %) basic ($\text{SiO}_2 = 46.1\text{-}50.9$ wt. %) type, which comprises mafic lamprophyres and pyroxenite xenoliths, and 2) a MgO-poor (1.5-6.4 wt. %) basic to intermediate ($\text{SiO}_2 = 48.2\text{-}61.2$ wt. %) type, mainly consisting of trachytes, felsic lamprophyres, and analcime-bearing tephriphonolites and phonotephrites. Rocks in both groups have variable but generally low Ni (10-115 ppm; mostly <60 ppm) and Cr (6-628 ppm; mostly <200 ppm), and display enrichment in light rare earth elements (REE) over heavy REEs (e.g., $\text{La/Yb} = 17\text{-}55$), with a large overlap between the MgO-rich and MgO-poor types. These features are coupled with enriched LILE/HFSE ratios (e.g., Ba/Nb mostly 16-116), negative Nb-Ta-Ti anomalies and positive Pb anomalies in primitive mantle-normalized diagrams (e.g., Nb/Nb^* mostly <0.7). Both high- and poor-MgO samples have strongly radiogenic ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratios (0.7078-0.7086), ($^{143}\text{Nd}/^{144}\text{Nd}$)_i below CHUR (0.51235-0.51244), and ϵHf_i ranging from -3.1 to -11.1‰, suggesting involvement of a lower continental crust component in their genesis. Their major oxide and trace element contents indicate a metasomatized subcontinental lithospheric mantle (SCLM), fluids and melts released by a subducted slab in the mantle wedge, as the magma source of the high-MgO and high-CaO rocks. This SCLM was evolved via mafic mineral fractionation and possible assimilation of lower crust, generating the higher SiO_2 and lower CaO and MgO-poor group. We suggest that small degree partial melting of the enriched mantle sources could have been triggered by Neo-Tethys slab roll-back during the Arabian-Iranian collision.

Keywords: ultrapotassic magmatic rocks; mantle peridotite; xenoliths; Eslamieh Peninsula; NW Iran

Introduction

Northwest Iran and East Anatolia regions form much of the Turkish-Iranian High Plateau, a key area for studying metasomatic processes and the modification of continental lithosphere during multi-stage subduction and collisional events. The petrological importance of this area is linked to its widespread igneous activity, with products showing highly variable chemical compositions and occasionally associated with the presence of mantle and crustal xenoliths (e.g., Moghadam et al., 2014; Neil et al., 2015; Sepidbar et al., 2021; Rabiee et al., 2022). Magma underplating and peridotite-melt reactions (Moghadam et al., 2014; Khedr et al., 2014) have been proposed as the two main mechanisms responsible for heterogeneous modification of the deep lithosphere.

The uplift of the Turkish-Iranian Plateau was one of the most important geodynamic events that occurred during the Cenozoic, closely related to the Arabian-Anatolian continental collision at ~25 Ma (Allen et al., 2013; Gusmeo et al., 2021). Post-collisional K-rich magmatic rocks, widely distributed across the Turkish-Iranian Plateau, have been the focus of much interest in recent years owing to their importance for understanding the nature of the upper mantle (Pang et al., 2013; Moghadam et al., 2014; Sepidbar et al., 2021; Rabiee et al., 2022). Deciphering the cause of igneous activity has implications for understanding the geodynamic setting of the region during the Cenozoic and for identifying the mechanisms of plateau uplift (e.g., Zhang et al., 2017).

Despite several studies, the petrogenesis and the identification of the processes responsible for the formation of post-collisional K-rich magmatic rocks remain highly controversial (e.g. Chung et al., 2009; Zhao et al., 2009; Guo et al., 2015; Lustrino et al., 2019, 2022). Three main hypotheses have been proposed in literature: (1) convective removal of the thickened lower part of the continental lithospheric mantle that led to passive asthenospheric upwelling during the extensional collapse that triggered igneous activity (Pang et al., 2013; Zhao et al., 2009); (2) intracontinental subduction of the continental lithosphere (Tapponnier et al., 2001), leading to a highly contaminated mantle source. Delamination of subducted continental materials may have played an essential role in the genesis of the ultrapotassic rocks (Zhao et al., 2009); and (3) rollback and break-off of a subducted slab of the continental lithosphere (Moghadam et al., 2018; Chen et al., 2015; Guo et al., 2015; Zhang et al., 2017). In this scenario, slab retreat and upper-plate extension, partial melting of the subducting slab and a fluxed high-temperature metasomatized mantle wedge would have led to eruption of the potassic magmas in a back-arc setting.

Post-collisional K-rich magmatic rocks in the Eslamieh (or Eslami or Eslamy) Peninsula, also known as the "Saray volcano" (Moghadam et al., 2014), formed alongside evolution of the eastern Turkish-Iranian Plateau. In this study, we investigate the genetic relationships between ultramafic xenoliths and host magmas in order to develop a model for generation of post-collisional K-rich magmas in this region. The involvement of enriched/depleted continental lithospheric mantle regions, recycled continental crust, subducted oceanic slabs, and delaminated and relaminated basaltic

lithologies have been suggested as potential mechanisms to generate the various types of potassic and ultrapotassic melts in active subduction and post-collisional tectonic settings (e.g., Lustrino et al., 2011; Förster et al., 2019; Bowman et al., 2021; Wang et al., 2021). The presence of potassic and ultrapotassic melts in the Eslamieh peninsula has been ascribed to small-degree partial melting of a subduction-metasomatized (subcontinental) lithospheric mantle source in a post-collisional setting (Moghadam et al., 2014). In this study, we propose that such mechanisms produced pyroxenites in the Eslamieh Peninsula, which then became entrained in the post-collisional K-rich magmas, as observed in other regions (e.g., Khedr et al., 2014).

We present new whole-rock major oxide and trace element data, Sr-Nd-Hf isotope data, and major oxide and trace element concentrations in minerals from K-rich magmatic rocks in the Eslamieh Peninsula, plus one zircon U-Pb age for a trachytic sample (Fig. 1). We also discuss the relationship between ultramafic xenoliths and their host magmas. We propose a subduction-metasomatized (subcontinental) lithospheric mantle source for these magmas, which subsequently evolved via fractionation and assimilation of lower crustal lithologies during ascent. The new mineral and whole-rock geochemical data for a suite of pyroxenite xenoliths constrain their origin in the deep lithosphere. The results of this study significantly improve our understanding of the processes involved in the lithospheric modification of the Turkish-Iranian Plateau.

2. Geological background

The Turkish-Iranian High Plateau is a composite fold-and-thrust belt formed by amalgamation of several crustal blocks (i.e., the Caucasus region of Georgia-Armenia-Azerbaijan, the eastern Pontides of Turkey, the NW Urumieh-Dokhtar magmatic belt, and Alborz of Iran) during Neo-Tethyan subduction, and the subsequent collision between Arabia and Eurasia (McQuarrie and van Hinsbergen, 2013; Fig. 1A). The collision is thought to have begun between the Late Eocene-Oligocene in Eastern Turkey, and during the Early Miocene in Iran (e.g., Dilek et al., 2010; Chiu et al., 2013). Continental collision led to the formation of the Turkish-Iranian High Plateau (Allen et al., 2013; Gusmeo et al., 2021) and diffuse pre-, syn-, and post-collisional magmatism with diverse geochemical signatures (Eyuboglu et al., 2013a, b).

In the eastern part of the Turkish-Iranian High Plateau, widespread Late Neoproterozoic-Early Cambrian (Cadomian) rocks crop out. Their large variations in initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and ϵNd_i values suggest an interaction of juvenile arc magmas with reworked Archean crust (Moghadam et al., 2017). This crystalline basement is overlain by Cambrian-Ordovician sedimentary rocks (the Soltanieh, Barut, and Zagun Formations), Carboniferous-Permian alkaline igneous rocks related to opening of the Paleo-Tethys Ocean, and Upper Jurassic-Late Cretaceous flysch and platform carbonates (Jamali et al., 2010). These latter are mostly covered by younger salty-clay deposits of the Urumieh Lake (Moine-Vaziri et al., 1991; Moghadam et al., 2014). Late Miocene shoshonitic and ultrapotassic

magmatic rocks, containing a variable cargo of mantle and crustal xenoliths, occur in the eastern parts of the Turkish-Iranian High Plateau (Moine-Vaziri et al., 1991; Khezerlou et al., 2017), along the Urumieh-Dokhtar magmatic belt (UDMB, and in the Eslamieh Peninsula (Moayyed et al., 2008; Hajlailou et al., 2009; Moghadam et al., 2014; Aghazadeh et al., 2015; Lustrino et al., 2019). Cenozoic igneous activity is recorded by several stages of magma generation with different geochemical signatures in the Late Eocene, Late Miocene, and Pliocene-Quaternary (Dilek et al., 2010; Eyuboglu et al., 2013a,b; Lustrino et al., 2021; Fedele et al., 2022; Rabiee et al., 2022). Intense and widespread subalkaline to alkaline igneous activity took place during the Late Cenozoic, and abundant mantle-derived xenoliths have been reported from several volcanic fields, including Eslamieh Peninsula and Marand (Khezerlou et al., 2017; Fig. 1A, B). The Eslamieh Peninsula magmatic rocks were mostly emplaced at 11-10 Ma (Pang et al., 2013; Aghazadeh et al., 2015), are characterised by alkaline to strongly alkaline compositions [essentially ultrapotassic to potassic (or shoshonitic)], and all carry subduction-related geochemical fingerprints.

2.1 Field observations of Eslamieh Peninsula volcanic rocks and related xenoliths

The Eslamieh Peninsula includes Late Miocene potassic (or shoshonitic) to ultrapotassic volcanic rocks and related xenoliths that are mainly exposed northeast of Urumieh city (Fig. 1B). The oldest basement rocks in the Eslamieh Peninsula include Late Neoproterozoic magmatic and metamorphic basement, overlain by late Miocene volcanic rocks – the focus of this study – that are covered by Quaternary salty-clay deposits of Lake Urumieh (Fig. 1C).

Volcanic rocks

The volcanic rocks of Eslamieh Peninsula outcrop over an area of $\sim 150 \text{ km}^2$ and have been subdivided into six major lithologies: (1) pyroxenite-glimmerite, nepheline-syenite and monzonite xenoliths hosted by the volcanic rocks; (2) potassic (or shoshonitic) to ultrapotassic lavas composed of trachyte (10.7 Ma; this study), basanite, leucite-tephrite, phonolite (11 Ma; Pang et al., 2013); (3) a thick pyroclastic sequence including rock fragments with variation in size between 4–5cm to even $>1\text{m}$; (4) Trachytic dikes injected in the previously emplaced volcanic (unit 2) and/or in pyroclastic (units 3); (5) a late stage unit associated with the emplacement of lahar deposits, and (6) dikes with mafic and felsic lamprophyric trachytic and leucitic compositions, which cut the lahar deposits near the Saray village (Fig. 1C). This study focuses on the pyroxenite and glimmeritic xenoliths (unit 1), trachyte and phonolites (unit 2), mafic and felsic lamprophyres (unit 6). All the investigated lavas (unit 2) are porphyritic with a fluidal/trachytic matrix, while the xenoliths show a granular to porphyritic textures. Porphyritic lavas, mainly occurring in the central part of the Eslamieh Peninsula (Fig. 1B, 2A) and contain plenty of pyroxenite and glimmerite xenoliths (Figs. 2-3) They are intruded

in the Early Miocene pyroclastic rocks and overlain with Late Miocene conglomerate type sediments with pebbles of older rock units and lahar deposits (Fig. 1C).

Two major types of lamprophyres have been identified in the Eslamieh Peninsula: (1) mafic lamprophyres, characterized by dark green/black colour, and (2) felsic lamprophyres with pinkish-grey in colour. Both occur as dykes, sills and plugs in the pyroclastic sequences (unit 3) and lava flows (unit 2; Fig. 2E), and both are characterized by abundant phlogopite phenocrysts up to 1 cm in size. The sills range from ~0.5 to ~3 m in thickness and radiate away from the centre of the volcano towards the shores of Lake Urumieh, whereas the dykes vertically crosscut the pyroclastic layers.

Ultramafic xenoliths

Two types of ultramafic xenoliths have been investigated in our study: (1) pyroxenites in trachytes, phonolite lavas, and felsic lamprophyric dikes; (2) glimmerites in trachytes and felsic lamprophyres. Both pyroxenite and glimmerite xenoliths have a size ranging from <1 cm up to 10 cm in diameter and show sharp contacts with the host magmas.

4. Analytical techniques

Major-element compositions of pyroxene and phlogopite from trachyte sample 20(H), pyroxene from phonolite samples 26(H) and 24-1(H), pyroxene and phlogopite from pyroxenite xenolith 1-1(X), and phlogopite from glimmerite xenoliths 7-4(X) and 14(X) were analysed with a JEOL JXA-8100 electron probe at the Institute of Geochemistry, Chinese Academy of Sciences, Guiyang, China, using a 15 keV accelerating voltage and a 10 nA beam current. Multiple grains of phlogopite and clinopyroxene were analysed from each host rock and xenolith, with a minimum of five analyses per mineral grain. A variety of natural and synthetic standards were used to calibrate the major components, some of which were also measured as unknowns to monitor data quality. The ZAF model (Armstrong, 1995) was applied for matrix correction. The two-sigma analytical precision of most elements (Si, Al, Ca, Fe, Mg, Na, Mn) analysed by EPMA is ~2%. Trace elements in clinopyroxene and phlogopite were analysed using laser ablation inductively coupled plasma mass spectrometer (LA-ICP-MS) at the China University of Geosciences, Beijing. The LA-ICP-MS system consists of a 193 nm pulsed ArF excimer laser coupled to an Agilent 7500 quadrupole ICP-MS. Key isotopes for each element were measured in peak-hopping mode. A spot size of 80 µm and a repetition rate of 8 Hz were used. The NIST612 glass was used as an external calibration standard and the isotope ⁴³Ca was used as an internal standard. BHOV-2 glass was measured as an unknown to verify data accuracy. The analytical precision of most elements analysed by LA-ICP-MS is ~5-10%.

Whole-rock geochemical analyses were performed on the freshest and most representative samples (trachytic and phonolitic lavas, mafic and felsic lamprophyric dikes, and pyroxenite xenoliths; Table 1) at the Institute of Geochemistry, Chinese Academy of Sciences, Guiyang, China, after crushing all the selected samples to powders of less than 200 mesh. Major element concentrations were

determined by X-ray fluorescence (XRF). The measurement procedure and data quality were monitored by simultaneous analyses of repeated samples (one in ten samples) and standard samples GBW07103, GBW07105, GBW07110, GBW07111, and GBW07112. The analytical uncertainties are generally <1% for most major elements. Trace element (including REE) concentrations were determined using an Agilent 7700e ICP-MS. The compositions of repeated samples (one in ten samples) and the reference materials AGV-2, BHVO-2, BCR-2, and RGM-2 were also analysed to monitor data quality. The analytical uncertainties are better than 5% for most trace elements. Detailed analytical procedures have been reported by Liu et al. (2008).

Whole-rock isotopic analyses, including Lu-Hf and Sm-Nd, were performed on the five pyroxenite xenoliths; 1-1(X), 1-2(X), 2-1(X), 2-2(X) and 3(X), two phonolites; 26(H) and 24-1(H), and two trachytes; 20(H) and 15(H) by multiple collector ICP MS (MC ICP MS) at the China University of Geosciences, Beijing. Detailed procedures and analytical errors have been reported by Lazarov et al. (2009) and Shu et al. (2013).

U-Pb dating and trace element analyses of zircon from one trachyte [sample 20(H)] were conducted simultaneously by LA-ICP-MS at the Institute of Geochemistry, Chinese Academy of Sciences, Guiyang, China. Laser sampling was performed using a GeoLas Pro and an Agilent 7700x ICP-MS was used to acquire ion-signal intensities. Helium was used as the carrier gas. Argon was the make-up gas and was mixed with the carrier gas via a T-connector before entering the ICP. Each analysis incorporated a background acquisition of approximately 10 s (gas blank) followed by 40 s data acquisition from the sample. An Agilent Chemstation was utilized to acquire each individual analysis. Off-line selection and integration of background and analyte signals, and time-drift correction and quantitative calibration for trace element analyses and U-Pb dating were performed using Glitter 4.4 software. Details of the instrumental conditions and data acquisition procedures are described by Li et al. (2015). Zircon 91500 was used as external standard for U-Pb geochronology. Time-dependent drifts of U-Th-Pb isotopic ratios were corrected using a linear interpolation. Uncertainty of preferred values for the external standard 91500 was propagated to the ultimate results of the samples. Concordia diagrams and weighted mean calculations were made using the Isoplot/Ex_version 3 software (Ludwig, 2003). Trace element compositions of zircons were calibrated against NIST610 standard combined with Si for internal standardization. The preferred values of element concentrations for the NIST610 reference glasses are from the GeoReM database (<http://georem.mpch-mainz.gwdg.de/>).

5. Analytical results

5.1 Petrography

We used mineral assemblages and petrography to divide our samples from the Eslamieh Peninsula into trachyte and phonolite lavas, mafic and felsic lamprophyric dikes, and xenoliths. All of the studied lavas are porphyritic with a fluidal/trachytic matrix, while the xenoliths show a granular texture. Porphyritic trachytic lavas contain ~10-15 vol. % phenocrysts of euhedral sanidine and clinopyroxene (Fig. 2D), with subordinate phlogopite (4-5 vol. %) and olivine (<2 vol. %), the latter showing reaction rims, especially in the most evolved compositions. Sanidine occurs as megacrysts with an average size of 3-4 cm (up to 10 cm), as well as forming in the groundmass, where it is aligned to form a trachytic texture (Fig. 2C). Secondary calcite is common in the groundmass of the trachytes, together with magnetite, apatite, titanite, and zircon.

Phonolites have porphyritic textures, with clinopyroxene phenocrysts (~10-15 vol. %) set in plagioclase, leucite, clinopyroxene, sanidine, and an apatite-rich groundmass. Two types of clinopyroxene crystals can be discriminated based on their size and colour: 1) euhedral to subhedral green clinopyroxene, and 2) large (>1 mm) subhedral clinopyroxene with pale-green or colourless cores and green rims. Leucite occurs as a groundmass phase (average size of 0.2-0.3 mm) and is generally analcimized.

The mafic and felsic lamprophyres are characterized by abundant phlogopite, with phenocrysts up to 1 cm in size. Typically, mafic lamprophyres contain phenocrysts of phlogopite (25-35 vol. %) and pyroxene (10-15 vol. %), and microlites of sanidine (5-10 vol. %), without any plagioclase (Fig. 2F-3A). These phenocrysts are surrounded by a groundmass with the same mineralogy plus apatite. These characteristics are typical of minettes (i.e., phlogopite-rich, plagioclase-poor/free lamprophyres). Phlogopite is the major phenocryst in mafic lamprophyres, with sizes up to 1 cm, followed by clinopyroxene, magnetite and apatite. Phlogopite phenocrysts also show clear signs of mechanical stress, including fractures and deformation features, such as kink-bands

The felsic lamprophyres have a porphyritic texture, with phlogopite (25-30 vol. %; up to 1 cm in size) and olivine (~5 vol. %; up to 1 cm in size) embedded in a grey to pinkish-grey groundmass with sanidine, phlogopite, magnetite and apatite plus secondary calcite (Fig. 3B, 3D). They carry abundant irregularly-shaped pyroxenites and glimmerite xenoliths up to 25-30 cm in length (Figs. 3B-D).

Ultramafic xenoliths

Pyroxenite xenoliths in trachytes show the same phases observed in the trachytic host magma, albeit with different modal abundances. The main component is anhedral poikilitic clinopyroxene (50-70 vol. %) with apatite inclusions (Fig. 3E-F), plus rare, small (<0.5 mm), and anhedral phlogopite (<5 vol. %) and quartz (<2 vol. %). Some clinopyroxene grains have clear cores and spongy rims with a width of ~0.5 mm.

Glimmerite xenoliths in felsic lamprophyre mainly consist of phlogopite (~50-80 vol. %) and clinopyroxene (~20-40 vol. %). Both are intergrown with subordinate euhedral apatite and magnetite crystals, or contain euhedral apatite inclusions. In addition, some phlogopite crystals contain carbonate blebs. Occasionally, spaces between phlogopite crystals are filled with interstitial glass.

5.1 Zircon U-Pb geochronology

Trachytic sample 20(H) contains zircon crystals without inherited cores. The grains are euhedral, prismatic in shape, and have average lengths of ~100-150 μm . The analysed zircons have high U (~780-2060 ppm) and Th (~200-750 ppm) contents, with relatively high Th/U (0.22-0.45) ratios (Supplementary Appendix 1), indicating a magmatic origin (Hoskin and Schaltegger, 2003). They yielded a U-Pb concordia age of 10.72 ± 0.18 Ma (Fig. 4b; Supplementary Appendix S1 and Fig. 4), which is interpreted as the crystallization age of the host magma. This age falls within the $^{40}\text{Ar}/^{39}\text{Ar}$ range reported in the literature (10.33-10.93 Ma; Pang et al., 2013; Moghadam et al., 2014; Supplementary Appendix S1).

5.2 Mineral compositions

EPMA and LA-ICP-MS analyses of mineral compositions mainly focused on clinopyroxenes and phlogopites in trachytic and phonolitic rocks, and pyroxenite and glimmerite xenoliths. The full set of data is reported in Supplementary Appendix S2a for clinopyroxene and S2b for phlogopite.

Clinopyroxenes are visibly zoned in back-scattered electron (BSE) images (Fig. 5). In trachytic samples, some clinopyroxenes show normal zoning, with dark cores and bright rims, whereas others are characterized by oscillatory zoning (Fig. 5A-1; Supplementary Appendix S2a). Inter-grain compositional variations are minor, with all analyses in the diopside field ($\text{Wo}_{46-49}\text{En}_{39-44}\text{Fs}_{7-11}$) and having Mg# [$\text{Mg\#} = \text{Mg}/(\text{Mg} + \text{Fe}^{2+})$] values ranging between 0.85 and 0.91. Dark core zones are characterized by higher MgO (14.9-15.6 wt. %) and Mg# (0.88-0.92) than those of bright zones, with MgO = 12.6-15.1 wt. % and Mg# = 0.85-0.88 (Supplementary Appendix S2). The higher Mg# compositions resemble those of clinopyroxenes from mantle peridotites (Hao et al., 2012; Khedr et al., 2014; Wang et al., 2014; Fig. 7A, C), whereas the lower Mg# clinopyroxenes are similar to those formed by melt-peridotite reactions (Fig. 7C; Bodinier et al., 2008).

Clinopyroxene in phonolites is augite-diopside characterized by normal zoning (Fig. 5B-1), with Mg-rich cores and Mg-poorer rims (Fig. 5B-2). They have a small compositional range ($\text{Wo}_{44-49}\text{En}_{40-48}\text{Fs}_{6-11}$), with minor acmite (<2.5 %) and kushiroite ($\text{Al}_2\text{O}_3 = 1.1-7.9$ wt. %; not shown; Supplementary Appendix S2) components similar to the clinopyroxenes in trachytes.

Clinopyroxene from pyroxenite xenoliths shows slightly oscillatory zoning, with compositions of $\text{Wo}_{45-47}\text{En}_{43-49}\text{Fs}_{4-9}$ range (Fig. 5C-1, C-2). Some crystals show complex sieve textures with resorbed

cores and bright rims with low MgO and high Al₂O₃ contents. No substantial differences were observed between clinopyroxenes in the xenoliths and those in the lavas, except that the kushiroite component tends to be lower in the former (Al₂O₃ = 1.0-3.9 wt. %; not shown; Supplementary Appendix S2).

Clinopyroxene in both the trachytic and phonolitic lavas has a Ni content (27-93 ppm) relatively similar to that in the xenoliths (40-129 ppm; Supplementary Appendix S2a). All pyroxene compositions show similar LREE enrichment over HREE (La_N/Yb_N = 3.1-5.7 and 3.3-8.1 for trachyte and phonolite, respectively), with slight enrichment of MREE over LREE patterns and nearly flat CI-normalised HREE profiles (Fig. 6). However, higher MgO contents mostly occurred in clinopyroxene from trachytic and phonolitic rocks (Supplementary Appendix S2a), whereas those from pyroxenite xenoliths (Supplementary Appendix S2b) have lower trace element contents (e.g., average Σ REE = ~98 ppm) than low-MgO zones (average Σ REE = 185 ppm). Clinopyroxenes have a clear depletion in highly incompatible trace elements (e.g., average Rb = 0.22 ppm; Ba = 0.69 ppm; Supplementary Appendix S2a) and a highly variable Cr content (~10-3390 ppm), with clinopyroxenes from the trachytic and phonolitic lavas showing the lowest values (~10-400 ppm; Supplementary Appendix S2a).

Phlogopite was analysed from trachytic lavas (Supplementary Appendix S2c) as well as pyroxenite and glimmerite xenoliths (Supplementary Appendix S2d). Phlogopite in trachytic lavas have TiO₂ in the 1.7-3.7 wt. % range and MgO-rich compositions (MgO = 18.8-25.1 wt. %; Mg# = 0.82-0.91). Compositions are similar to phlogopite in peridotite xenoliths hosted in alkali-rich magmas worldwide (Figs. 7E; Lustrino et al., 1999; Pearson et al., 2003). The phlogopite found in pyroxenite xenoliths show lower MgO (15.2-17.9 wt. %) and Mg# (0.68-0.76), coupled with higher TiO₂ (3.9-5.2 wt. %) than those in trachytes (Supplementary Appendix S2c).

In contrast to major oxide composition, phlogopite from host rocks and xenoliths is indistinguishable in terms of trace elements, having relatively low REE (0.06-0.7 ppm and 0.1-0.7, respectively), Sr in the ranges 20-287 and 98-215 ppm), Rb in the ranges 131-565 and 153-409 ppm, and Li in ranges 1.4-8.2 and 1.5-9.0 ppm, as well as high Ba in range of 4585-41771 and 5466-48405 ppm), respectively (Supplementary Appendix S2b).

5.3 Whole-rock chemistry

Whole-rock data for Eslamieh Peninsula rocks are reported in Table 1, with additional literature analyses from the same area provided in Supplementary Appendix S3. The investigated rocks plot in the ultrapotassic field in the K₂O vs. Na₂O diagram, being characterized mostly by K₂O >3 wt. % and K₂O/Na₂O >2 (Fig. 8A; Supplementary Appendix 3). Some literature analyses of Eslamieh Peninsula volcanic rocks plot also in the potassic and transitional fields in the same classification diagram (Fig. 8A). The Saray volcanic rocks and pyroxenite xenoliths were classified using the total alkalis

(K₂O+Na₂O wt.%) vs. SiO₂ of Le Bas et al. (1986) (Fig. 8b) and Zr/Ti vs. Nb/Y diagrams (Fig. 8C). The rocks fall in a wide range represented by tephrite to phonolite, trachyte and trachy-andesite fields, in agreement with their petrographic features. The mafic lamprophyres cluster in the basanite and phonotephrite fields and pyroxenite xenoliths plot in the alkali basalt, trachybasalt, and basaltic trachyandesite fields in the TAS diagram (Fig. 8B). The two samples with intermediate compositions plot in the phonotephritic and tephriphonolitic fields (Fig. 8B). We refer to these samples as “phonolites”, although they do not plot in the TAS phonolite field. On a Zr/TiO₂ vs. Nb/Y diagram (Pearce, 1996), most of the samples cluster in a relatively restricted area between the trachyte-trachyandesite fields (Fig. 8C).

Albeit not representing real magma compositions, the pyroxenite xenolith compositions are shown in the same figure, where they straddle the potassic/ultrapotassic fields. Measured loss on ignition (LOI) values are low in the trachytic samples (0.8-1.6 wt. %) to moderately high (1.2-4.6 wt. %) in the mafic lamprophyres and pyroxenite xenoliths, to very high in the felsic lamprophyres and phonolitic rocks (5.2-6.6 wt. %). The xenolith and mafic lamprophyre are characterized by higher MgO, Ni and Cr, lower SiO₂ (46.0-50.7 wt. %) and generally lower Na₂O (excluding one xenolith) contents, coupled with higher Mg# [0.65-0.94; Mg# = Mg/(Mg + Fe²⁺), assuming Fe₂O₃/(FeO + Fe₂O₃) = 0.15], hereafter introduce as high-MgO subgroup compared with the trachyte, phonolite and felsic lamprophyre, known as low-MgO subgroup (Supplementary Appendix S3).

Our analyzed rock compositions almost completely overlap the compositions of Eslamieh samples reported in the literature, and show either roughly positive (TiO₂, Fe₂O_{3tot}, CaO, and P₂O₅) or negative correlations (SiO₂, Al₂O₃, and alkalis) with MgO (Fig. 9). The pyroxenite xenolith samples generally do not follow the trends shown by the lava samples (Fig. 9).

Primitive mantle-normalised diagrams show spiky patterns for all investigated samples, with common features represented by troughs at Nb-Ti, coupled with positive peaks at Pb and K, and a general enrichment of LILE compared to HFSE (Fig. 10). However, low-MgO samples, except for felsic lamprophyres, are more enriched in Pb with respect to high-MgO one (Fig. 10). All the rocks have very low U contents, with the strongest depletion observed in some trachytes and pyroxenite xenoliths (Fig. 10).

No substantial differences are observed in CI chondrite-normalized REE patterns for mafic and felsic lamprophyres (Fig. 10)), with the former showing slightly higher absolute concentrations than the latter. Overall, a large inter-elemental fractionation overlap is observed, with La_N/Yb_N ranging from 25 to 31 and from 23 to 51 for mafic and felsic lamprophyres, respectively. Both lamprophyre types are characterised by strong negative Eu anomalies (Eu/Eu* = 0.42-0.76 and 0.58-0.89 for the mafic and felsic variants, respectively). The trachytic rock group shows complete compositional overlap with the lamprophyres (La_N/Yb_N = 20-37; Eu/Eu* = 0.36-0.80; Fig. 10), whereas the pyroxenite xenoliths (La_N/Yb_N = 17-20; Eu/Eu* = 0.72-0.91) and the phonolites (La_N/Yb_N = 12-16;

Eu/Eu* = 0.77-0.84) tend to show lower REE enrichment, lower LREE/HREE fractionation and smaller Eu negative anomalies (Fig. 10).

5.4 Sr-Nd-Hf isotopes

Whole-rock Sr-Nd-Hf isotopic data for two trachytic rocks, two phonolitic rocks, and five pyroxenite xenoliths are presented in Supplementary Appendixes 3, and are illustrated in Fig. 11. Eslamieh lavas have strongly radiogenic ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ratios (0.70824-0.70863) and strongly unradiogenic Nd ($^{143}\text{Nd}/^{144}\text{Nd}$)_i (0.51235-0.51240) ratios. These compositions trend towards the EM2 (enriched mantle type-2) OIB component, which is ultimately related to interactions with subducted lithologies in mantle sources (e.g., Workman et al., 2004; Lustrino and Anderson, 2015) with affinity to the other subduction-related potassic and ultrapotassic volcanic rocks of the Alps and the Betics in SE Spain (Fig. 11A). One of the two analysed xenoliths has Sr-Nd isotopic compositions indistinguishable from those of the lavas, while the second one is characterised by less radiogenic Sr [$(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.70780$] and more radiogenic Nd [$(^{143}\text{Nd}/^{144}\text{Nd})_i = 0.51245$]. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios are relatively low, clustering around 0.28246-0.28268 for the xenoliths and 0.28257-0.28266 for the lavas, values that correlate well with Nd isotopic ratios. The ϵHf_i of the xenoliths (from -3.1 to -11.1) and related host rocks (from -3.9 to -7.2) allow us to infer the presence of crustal components either in the mantle source, or entrained during melt ascent and emplacement (Fig. 11B).

6. Discussion

The origin and evolution of magmas emplaced along active margins comprise a series of processes that include (e.g., Eyuboglu, 2013; Fedele et al., 2022): (1) metasomatism of the mantle by melt and/or fluids released by the subducting slab (e.g., Hawkesworth et al., 1997a, b; Khedr and Arai, 2009; Khedr et al., 2010); (2) variable partial melting of the metasomatized mantle (e.g., Pearce and Parkinson, 1993); (3) magma mixing and/or assimilation plus fractional crystallization (AFC; e.g., DePaolo, 1981), and/or (4) combinations of these processes (Ersoy et al., 2010). Here we discuss the role of each process trying to decipher the origin of the pyroxenite xenoliths, to understand the petrogenesis of Eslamieh magmas and to put these results in a geodynamic framework.

6.1. Origin of the pyroxenite xenoliths

Pyroxenite comprises a volumetrically minor portion of the upper mantle, although it can significantly influence the composition of peridotite partial melts with which it interacts. These compositions range from orthopyroxenites through websterites to clinopyroxenites, with or without olivine, spinel, and garnet as minor constituents (Downes, 2007). Previous studies have shown that pyroxenite xenoliths have diverse origins. They can form as (i) high-pressure cumulates from mantle-derived melts (Xu, 2002); or (ii) gabbroic cumulates from mantle-derived melts, which were transformed to pyroxenites

during subsequent subduction (Yu et al., 2010), or (iii) can result from melt and/or fluid-peridotite reactions (Hu et al., 2016). Pyroxenites usually preserve a more extensive reaction history compared to the possibly associated lherzolites, and thus they are particularly useful for investigating metasomatic processes, the recycling of crustal components, and the thermal and chemical modification history of upper mantle (Xu, 2002). Pyroxenite xenoliths analysed in this study are characterised by relatively low SiO₂ (48.8-50.9 wt. %) and CaO (13.3-17.0 wt. %), coupled with high MgO (8.7-13.9 wt. %), and Mg# (0.84-0.94), features that are inconsistent with their origin by fractional crystallization of any primary magma. In addition, pyroxenes from pyroxenite xenoliths from Eslamieh Peninsula have low Al₂O₃ contents (3.6-5.6 wt. %), and their chondrite-normalized REE patterns are characterized by low Eu negative anomalies (Fig. 6E). These compositions rule out the formation of the pyroxenite xenoliths as recycled oceanic lower crust or high-P/T metamorphosed gabbroic cumulates, which instead have high Al₂O₃ contents (>12 wt. %) and positive Eu anomalies (Yu et al., 2010). The ⁸⁷Sr/⁸⁶Sr higher than Bulk Silicate Earth (BSE) estimate, coupled with ¹⁴³Nd/¹⁴⁴Nd lower than Chondritic Uniform Reservoir estimates (ChUR) concord in excluding a MOR-related protolith for these compositions. Clinopyroxenes in the Eslamieh pyroxenites have Mg# contents similar to typical mantle peridotite xenoliths (Fig. 7), but the LREE enrichment of both whole rock samples and clinopyroxene crystals (especially in the rims; Fig. 9) is consistent with the pyroxenites being the products of melt-peridotite reactions (Fig. 7D; Bodinier et al., 2008; Chin et al., 2020).

6.2 Petrogenesis of Eslamieh magmas

6.2.1 Mantle source regions and subduction-related metasomatism

Whole-rock major oxides and trace elements can be used to trace the presence of different lithologies (e.g., mantle pyroxenite/peridotite) in a magma source region (Herzberg and Asimow, 2008; Guo et al., 2015). Peridotite partial melts are characterized by relatively low SiO₂ coupled with high MgO and CaO (Sobolev et al., 2007; Guo et al., 2015) characteristics observed in the mafic lamprophyres and pyroxenite xenoliths from Eslamieh Peninsula. These also show relatively low Ni, Cr, SiO₂ (46.0-50.1 wt. %), Na₂O contents and Ni/MgO ratios, and relatively high Mg# (0.65-0.69). By contrast, the volcanic rocks belonging to the MgO-poor group are very different from melts in equilibrium with the mantle, as they are depleted in MgO, Ni, and Cr, and enriched in SiO₂. Given the origin of the studied pyroxenites (see section 6.1), we also consider the mafic lamprophyres to constrain the nature of the mantle source regions of Eslamieh Peninsula magmas.

Relatively fluid-immobile incompatible elements, such as HFSE and HREE, can track enrichment or depletion processes in the source regions of magmas (Khedr and Arai, 2009; Khedr et al., 2010, 2014; Hawkesworth et al., 1997a, b; Elliott et al., 2003). In a Th/Yb vs. Nb/Yb diagram, the high-MgO samples plot above and between E-MORB and OIB composition fields, towards strongly enriched Th/

Yb ratios (Fig. 13A, B). To constrain the origin of such enrichment, the samples are plotted in the Th/Nd vs. Ba/La diagram (Wang et al., 2004). High Ba/La (>10) ratios indicate metasomatism by fluids released from subducted slabs (Hart, 1988; Wang et al., 2004), whereas high Th/Nd ratios (>1) indicate metasomatism by slab-derived melts (Hart, 1988; Wang et al., 2004). The high Ba/La ratios (26-49) coupled with low Th/Nd (<1) of Eslamieh mafic lamprophyres (Fig. 13C) strongly suggest the involvement of slab-derived fluids rather than melts.

Only the most differentiated Eslamieh ultrapotassic rocks fall within the field of the Tethyan realm lamproites in the Th/La vs. Sm/La diagram (Fig. 13D; Tommasini et al., 2011; Prelević et al., 2013; Guo et al., 2015), with high Th/La and Sm/La ratios up to 1.1 and 0.23, respectively (Fig. 12D). By contrast, the mafic lamprophyres and the pyroxenite xenoliths show lower Th/La and higher Sm/La, clustering closer to the fields of “normal” arc magmas. The Th/La enrichment identified by Tommasini et al. (2011) in Tethyan-realm lamproites, is reported, however, also in non-lamproitic compositions, and consequentially it cannot be considered a peculiarity of lamproitic compositions (e.g., Lustrino et al., 2016; Wang et al., 2021). These geochemical characteristics require the involvement of specific components (such as epidote or lawsonite) that can host high trace elements (REE, Th-Pb) contents (Prelević et al., 2013; Wang et al., 2021). Lawsonite is an abundant hydrous mineral that forms in metamorphosed oceanic crust, sediments, and metasomatic rocks at depths of ~ 45 to 300 km in cold subduction zones. Experimental (Hermann et al., 2013) and petrological modelling (Palin and White, 2016) studies have demonstrated that eclogite-facies metamorphism of subducted slab may be accompanied with breakdown of lawsonite with increasing temperature; therefore partial melting of mantle metasomatised with such hydrous and incompatible-rich phases (e.g. Foley, 1992; Mitchell, 1995; Kirchenbaur et al., 2012), can produce K-rich magmas with high Th/La and Sm/La ratios (Fig. 13D). The Rb/Sr and Ba/Rb ratios, and relatively high K_2O contents of the both more primitive rocks (Fig. 13E), may also suggest the presence of phlogopite participating to partial melting (Fig. 13E; Furman and Graham, 1999), which is further supported by presence of phlogopite in the pyroxenite xenoliths and trachytic hosts in the Eslamieh volcanic field.

Fractionation of REE (e.g., La/Yb, Sm/Yb or Dy/Yb) is affected by the depth and degree of melting in the mantle source regions (Shaw et al. 2003; Oyan et al. 2017). Decompression melting of metasomatized lithospheric lherzolite with minor phlogopite and pargasite has been invoked for the formation of primary high-K and high-MgO magmas by dehydration melting at ~ 1 GPa, 1050-1150 °C (Conceição and Green, 2004). Our whole rock geochemistry data [Th/Yb: 1.6-21.5, Nb/Yb: 2.7-30.5, and high Ba/La: >10 , La/Yb: 23.6-45.5 and Dy/Yb: 2.9-3.5] reveal that the source regions of Late Miocene, more primitive, high-MgO rocks were affected by somewhat different degrees of contamination by slab-fluid components and degrees of partial melting. A melt curve model was calculated based for La/Yb and Dy/Yb ratios using non-modal batch melts of garnet and spinel lherzolites for two different enriched mantle source compositions: S1 (from Saccani et al., 2013a) and

S2 (from Saccani et al., 2013b; Fig. 14). The Dy/Yb (3.0-3.5) ratios of the mafic lamprophyres, coupled with their low HREE (42-46) contents support the derivation from a garnet-enriched peridotite (e.g., McKenzie and O’Nions, 1991; Table 1). We also note that the La/Yb-Dy/Yb systematics of these rocks are compatible with ~1-2% partial melting of a lherzolite with garnet/spinel ratio >2 (Fig. 14).

6.2.2 Rock associations and magmatic evolution

Both of the Eslamieh high- and low-MgO rocks mainly record potassic (or shoshonitic) to ultrapotassic signatures and show geochemical characteristics of arc magmatic rocks. The geochemical signatures of trachyte, phonolite, felsic lamprophyre, mafic lamprophyre and pyroxenite including depletion in Nb, Ti and enrichment in LILEs and LREEs/ HREEs, are similar to the geochemical characteristics of continental arc magmatic rocks (Ducea et al., 2010), suggesting, they may be expressions of the same magma. The Eslamieh Peninsula trachyte, phonolite and felsic lamprophyres have low MgO contents (as little as 1.5 wt. %), low Mg# (<0.5), and low Ni (10-37 ppm) and Co (9-61 ppm) concentrations. We interpret this to be a result of enhanced and prolonged fractional crystallization. Major oxide (e.g. SiO₂, Al₂O₃, TiO₂, Na₂O+K₂O, P₂O₅, and CaO) variation vs. MgO could provide evidence for fractionation trends (Fig. 9). In particular, the correlation between Fe₂O₃T, CaO, and TiO₂ vs. MgO (Fig. 9) indicates a trend controlled by pyroxene fractionation, with a minor contribution by Fe-Ti opaque minerals (Fig. 9; Supplementary Figs. S1 and S2), albeit oscillatory zoning in clinopyroxene and Nd-Sr-Hf isotopic ratios could imply other mechanisms, such as AFC and/or magma chamber replenishment. However, petrographic (e.g. the presence of complex zoning of the clinopyroxenes) and geochemical features of the lavas (e.g., high abundance of LREE and LILE) and high ⁸⁷Sr/⁸⁶Sr values cannot be explained either by FC nor by AFC processes alone, at least when assuming potential contaminants to be local crystalline basement rocks, such as the Zanjan-Takab complex in NW Iran (Moghadam et al., 2018). The Sr-Nd isotopic ratios reported in this study can be explained assuming AFC processes between a mafic magma (represented by sample E11-29 from Eslamieh Peninsula; Moghadam et al., 2014) and a specific contaminant, such as the Cadomian granitic gneiss sample ZN14-4 of the Zanjan-Takab complex in NW Iran, which is similar in age and composition to the basement rocks of Eslamieh Peninsula (Moghadam et al., 2018) (Supplementary Fig. S3). It should be mentioned that since our primitive samples are characterized by strongly radiogenic (⁸⁷Sr/⁸⁶Sr)_i ratios (0.7078-0.7086), (¹⁴³Nd/¹⁴⁴Nd)_i below CHUR (0.51235-0.51244), and εHf_i ranging from -3.1 to -11.1‰, we used the less radiogenic sample E11-29 (Moghadam et al., 2014) as a better representation of an end member for melt derived from the mantle. We first considered an average fractionating composition of feldspar (65%) + clinopyroxene (15%) + amphibole (10%) + mica (10%) following Ersoy et al. (2012a) (Supplementary Fig. S3), as these considerations also fit well with our observations in the field showing clinopyroxene–phlogopite–feldspar (±olivine) in the rocks. However, in this case, the

assimilation/fractional crystallisation ratio would require up to 90% removal of crystallized phases, which is incompatible with the whole-rock isotopic signatures [$(^{87}\text{Sr}/^{86}\text{Sr})_i$ ratios: 0.7078-0.7086, $(^{143}\text{Nd}/^{144}\text{Nd})_i$: 0.51235-0.51244, and ϵHf_i ranging from -3.1 to -11.1‰] of the investigated rocks. In addition, the presence of ultramafic dense xenoliths in evolved rocks is inconsistent with prolonged stagnation at shallow crustal depths.

Instead, we considered other causes of magmatic evolution, including mantle source heterogeneity due to metasomatism driven by sediment fluids/melts released from subducting slabs (e.g., Ersoy et al., 2008; Conticelli et al., 2009). Field observations (e.g., Moayyed et al., 2008; Moghadam et al., 2014; this study) and radiometric ages of the phonolites ($\sim 10.7 \pm 0.2$ Ma; 11.0 ± 0.1 Ma; Pang et al. 2013), trachytes (Ar-Ar ages of mica: 10.32 ± 0.17 and 10.68 ± 0.25 Ma; Moghadam et al. 2014, U-Pb ages of zircon: 10.7 ± 0.5 ; this study), and felsic lamprophyres (Ar-Ar ages of mica: 10.35 ± 0.02 Ma; Aghazadeh et al. 2015) indicate that the foid-bearing volcanic rocks are slightly older than the lamprophyres and trachytes.

The co-occurrence of mafic and felsic lamprophyres is an unusual feature of the Eslamieh Peninsula, although such a pairing has also been reported in Buell Park, Arizona by Roden and Smith (1971). These authors suggest that both compositions formed due to crystal fractionation or different degrees of partial melting of the same source rock. Numerous studies point to the association of lamprophyres with alkaline rocks, such as trachytes and syenites (Nardi et al. 2012).

As shown previously, both the Eslamieh felsic and mafic lamprophyres have similar geochemical signatures, suggesting the same source for their genesis. The felsic lamprophyres show nearly complete major oxide and trace element overlap with the trachytic rocks and, compared to the mafic variants, they show higher SiO_2 and Al_2O_3 , but lower MgO , CaO , TiO_2 , P_2O_5 , $\text{Fe}_2\text{O}_{3\text{tot}}$ and MnO (Fig. 9) with respect to mafic lamprophyres.

Such an association may alternatively originate due to mixing between mafic lamprophyric and trachytic melts. This process is compatible with chemical variations from rim to core in phenocrysts following the recharge of a chemically different (i.e. more mafic) magma, and the plots of the felsic lamprophyre between mafic lamprophyre and trachyte in Harker-like diagrams (Fig. 9). Abundant K-feldspar (sanidine) megacrysts (up to 10 cm) in the trachyte samples also clearly indicate that they are broken native phenocrysts that crystallized from the magma under disequilibrium conditions, whereas K-feldspar in felsic lamprophyres forms the main groundmass phase. This is consistent with the general statement by Chapman and Powell (1976) who consider K-feldspar megacrysts in lamprophyres to be xenocrystic in origin.

7. Geodynamic implications

Potassic and ultrapotassic igneous rocks form in a wide range of tectonic settings, such as active continental margins, post-collisional, and within-plate settings (e.g., Lustrino et al., 2019). Collision between the Arabian and Iranian plates initiated at ~ 25 Ma (Moghadam and Stern, 2011), implying

that the late Cenozoic magmatism in the eastern parts of the Turkish-Iranian High Plateau and northwestern part of the UDMB occurred in a post-collisional setting. Consequently, the most likely tectonic setting is one where subduction-related metasomatized sub-continental lithospheric mantle experiencing partial melting (e.g., Prelević et al., 2012). Moghadam et al. (2014) suggested that the Late Miocene high-K magmas were derived from a small degree of partial melting of subduction-metasomatized (subcontinental) lithospheric mantle source in a post-collisional setting. Abundant pyroxenite xenoliths in the Eslamieh Peninsula confirm that mantle refertilization occurred beneath this area. It is clear from the trace and rare element distribution in minerals that the lithospheric mantle was modified by melt-rock interactions (melt/fluid metasomatism; e.g., Wang et al., 2019; Yang et al., 2018; Zhao et al., 2015) as indicated by high- and low-MgO pyroxene with different REE distribution (Fig. 6).

Based on whole-rock compositions (including Sr-Nd-Hf isotopic ratios), we propose that ultrapotassic magmas of the Eslamieh Peninsula mainly originated from partial melting of a metasomatised peridotitic source. Our conclusion agrees with that of Moghadam et al. (2014), who proposed that these magmas record the addition of melts and fluids released from the Tethyan lithospheric slab into the supra-subduction zone mantle wedge. High Ba/La (>10) ratios and negative Eu/Eu* signature in lack of plagioclase confirm metasomatism by fluids and melts released from subducted slabs (Hart, 1988; Wang et al., 2004). The same feature is observed in other plagioclase-poor to plagioclase-free ultrapotassic rocks in the central-western Mediterranean, interpreted with the presence of a non-carbonatitic sedimentary component (characterized by negative Eu/Eu* values) in their mantle sources (e.g., Conticelli et al., 2009; Lustrino et al., 2011). We disagree, however, with Moghadam et al. (2014), who proposed that the negative Eu/Eu* values of the Saray potassic and ultrapotassic rocks were derived from fractional crystallization of feldspars. Indeed, these authors suggest the presence of a parental melt that crystallized plagioclase (and/or amphibole) before producing the derivative magmas sampled in the Eslamieh Peninsula (Moghadam et al., 2014). The authors modelled the geochemical evolution of Eslamieh rocks by assuming an AFC-type process that involved fractionation of a plagioclase-rich (40-65%) assemblage, whereas our model suggests feldspar and phlogopite fractionation during the AFC process (Supplementary Fig. S3).

We propose that NE-directed underthrusting of the subducted slab released considerable volumes of melts and fluids that would have heterogeneously infiltrated the peridotitic matrix in the mantle wedge (Fig. 15). The continental collision between Arabia and Iran caused break-off of the oceanic slab, followed by passive upwelling of the sub-lithospheric mantle (Verdel et al., 2011). Decompression melting of the metasomatised lithospheric mantle with minor AFC process has been invoked for the formation of primitive high-K and high-MgO magmas by dehydration melting at ~ 1 GPa, 1050-1150 °C (Conceição and Green, 2004; Agard et al., 2005; Zhao et al., 2009; Fig. 15A). These processes could have initiated partial melting of the metasomatised mantle source, infiltrated heterogeneously with melt or fluid released by previous subducted slab in the mantle wedge, giving

rise to production of high-MgO ultrapotassic magmas. Low-degree partial melting of the metasomatized mantle and more AFC processes led to the formation of high-K and low MgO magmas enriched in incompatible trace elements (Fig. 15B). These evolved K-rich magmas would have rapidly upwelled and ascended towards the Earth's surface, entraining mantle xenoliths which are now exposed in the Eslamieh volcanic field lavas. Underplating of such mantle-derived magmas at the base of the lower crust has been widely recognized as an important process in crust-mantle evolution (Rudnick, 1995; Zheng et al., 2009b). This facilitates vertical crustal accretion and the formation of the crust-mantle transitional zone, which is a significant mechanism for the chemical evolution of the crust-mantle boundary (Fan et al., 2005).

8. Conclusions

- (1) The Eslamieh Peninsula hosts alkaline volcanic rocks with variable K₂O enrichment over Na₂O, that have compositions ranging from potassic to ultrapotassic, and basic to intermediate in terms of SiO₂ content (mafic to felsic lamprophyres, up to trachytes and phonolites). These lithologies are commonly associated with ultramafic (mantle) xenoliths. The xenoliths investigated in the present study are pyroxenites and glimmerites.
- (2) The rocks investigated in this study can be divided into two types based on their compositions: a basic (SiO₂ = 46.1-50.9 wt. %), high-MgO (8.0-13.9 wt. %) type, and a mafic to intermediate (SiO₂ = 48.2-51.4 wt. %), low-MgO (4.5-6.4 wt. %) (1.5-6.4 wt. %) type.
- (3) The trachyte and phonolites contain zoned clinopyroxene phenocrysts with MgO-rich and LREE-poor cores and MgO-poor and LREE-rich rims.
- (4) The low MgO volcanic rocks were produced by low degree partial melting (1-2%) of metasomatized mantle sources and evolved during AFC process. Variations in Sr-Nd-Hf isotopes are ascribed to source heterogeneities, as changes in isotopic ratios are unrelated to major oxide variations.

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Figure captions

Fig. 1. (A) Geological map of Iran showing Cenozoic magmatic rocks and the position of the Eslamieh Peninsula. (B) Geological map of the Eslamieh Peninsula and location of the studied rocks.

Fig. 2. Trachytic/trachyandesitic lavas located in the central part of the Eslamieh Peninsula; (B) sharp contact between trachytic host and pyroxenite xenoliths; (C) aligned sanidine crystals within lavas showing a trachytic texture; (D) euhedral clinopyroxene phenocrysts with light to dark green zonation; (E) individual dykes in lava flows; and (F) phlogopite grains in phlogopite-rich, plagioclase-poor/free lamprophyres.

Fig. 3. (A) Rounded phlogopite and pyroxene phenocrysts in a lamprophyric dyke; (B) phlogopite crystals up to 1 cm in size embedded in a grey to pinkish-grey groundmass of felsic lamprophyres; (C) and (D) irregularly-shaped xenoliths up to 25-30 cm long, consisting of pyroxenites and glimmerites in the felsic lamprophyres; (E) and (F) pyroxenite xenoliths containing anhedral pyroxene crystals with quartz and apatite inclusions.

Fig. 4. (A) Inverse and (B) weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age plots for zircons from trachytic samples from Eslamieh Peninsula.

Fig. 5. BSE-SEM images of (A-1) Trachytic host, (B-1) phonolitic host, and (C-1) pyroxenite xenolith showing; (A-2) oscillatory variation in Mg composition from cores to rims in the trachytic samples;

(B-2) Mg-rich cores and Mg-poor rims of pyroxenes in phonolitic samples, and clear oscillatory variation in compositions of pyroxene within the pyroxenite xenoliths.

Fig. 6. Chondrite-normalized rare earth and trace element diagrams for clinopyroxene in trachytes (A-B), phonolite (C-D), and pyroxenite xenolith (E-F). Normalization factors are from Sun and McDonough (1989).

Fig. 7. Chemical compositions of pyroxene and phlogopite phenocryst from ultrapotassic magmatic rocks in the Eslamieh Peninsula; (A-B) Pyroxene nomenclature based on Wo-En-Fs endmembers for volcanic host and pyroxenite xenoliths; (C-D) Al_2O_3 vs Mg# diagrams for clinopyroxene in volcanic host rocks and xenoliths; (F) (b) Phlogopite grains from K-rich magmatic rocks and xenoliths plotted on Al-Mg-Fe diagrams after Sheppard and Taylor (1992). Light yellow and turquoise areas represent the compositional ranges for phlogopite from shoshonitic lamprophyres and lamproites, respectively (after Sheppard and Taylor, 1992).

Fig. 8. K_2O vs. Na_2O (A), Alkali vs SiO_2 (B) Zr/TiO_2 vs Nb/Y (c), and Th vs Co (D) plots for the classification of potassic and ultrapotassic magmatic rocks from the Eslamieh Peninsula.

Fig. 9. Selected major element oxide (wt. %) vs MgO (wt. %) contents illustrating the compositional range of ultrapotassic magmatic rocks in the Eslamieh Peninsula. All the major element data were recalculated to 100 wt. % on a volatile-free basis. FeO* denotes total iron reported as FeO. Data sources and symbols are as in Fig. 8.

Fig. 10. CI chondrite-normalized REE-normalised (King et al., 2020) and primitive mantle-normalised (Lyubetskaya and Korenaga, 2007) diagrams for the ultrapotassic magmatic rocks: trachyte (A-B), phonolite (C-D), mafic and felsic lamprophyres (E-H), and xenoliths (I-J). Normalization factors are from Sun and McDonough (1989).

Fig. 11. (A) $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and (B) ϵ_{Hf} vs age (Ma) diagrams for the Saray high-K rocks compared with the mantle components HIMU, EM1, EM2, DMM (after Zindler and Hart, 1986); FOZO1 (after Hart et al., 1992) and ITEM (after Bell et al., 2004); present-day bulk Earth and CHUR (chondritic uniform reservoir) are also shown. Most of the data cluster close to the EM2 domain. Data for the Eastern and Western Alps lamprophyres are from Bell et al. (2013).

Fig. 12. (A) Plot of MgO vs CaO for the ultrapotassic magmatic rocks in the Eslamieh Peninsula. The CaO contents for peridotite partial melts were modelled by Herzberg (2011). (b) Plot of Ni vs MgO. Broken red lines separating peridotite- and pyroxenite-source primary magmas are after Herzberg and Asimow (2008) and Guo et al. (2015).

Fig. 13. (A-B) Th/Yb vs Nb/Yb diagram; (C) Th/Nd vs Ba/La; (D) Th/La vs Sm/La; and (E) whole-rock Rb/Sr vs Ba/Rb. Phlogopite and amphibole on the arrows refer to the phases that occur as

residues in the mantle source region. Petrological groups and references: BABB, back-arc basalts (Pearce et al., 2005; Buchs et al., 2013); FAB, forearc basalts (Reagan et al., 2010; Ishizuka et al., 2011); tholeiite-OIB, ocean island basalts (Willbold and Stracke, 2006; Buchs et al., 2013); E-MORB, enriched mid-ocean ridge basalts (after Jenner and O'Neill, 2012; Azizi et al., 2018a); the 85% probability contour of island arc tholeiite (IAT) compositions is after Hastie et al. (2007). The orange field in (D) shows the compositional variation of arc magmatic rocks related to slab-derived mantle metasomatism (Plank, 2005; Plank et al., 2007); the blue field covers the compositional trend of Tethyan-realm lamproites (Tommasini et al., 2011 and Prelević et al., 2013). Compositions of upper crust (UC) and lower crust (LC) are from Rudnick and Gao (2003). Global subducting sediment (GLOSS) is according to Plank and Langmuir. Data sources and symbols are as Figure 8.

Fig. 14. Melt curves for S1 and S2 mantle in both garnet- and spinel-facies; Arrays representing mixing between various proportions of melt fractions from garnet-facies mantle and melt fractions from spinel-facies mantle are also shown. Melt curve models based on La/Yb vs. Dy/Yb. Melt curves are calculated using non modal, batch melts of garnet and spinel lherzolites for two different enriched mantle source compositions: S1 (from Saccani et al., 2013a) and S2 (from Saccani et al., 2013b). Garnet lherzolite mode is: 0.598 ol, 0.211 opx; 0.076 cpx, and 0.115 gt that melts in proportions of 0.04 ol, - 0.19 opx, 1.04 cpx, and 0.11 gt. Spinel lherzolite mode is: 0.578 ol, 0.270 opx, 0.119 cpx, and 0.033 sp. that melts in proportions of - 0.06 ol, 0.28 opx, 0.67 cpx, and 0.11 sp. Mantle modes and melting proportions are from Kinzler (1997). Distribution coefficients are from Irving and Frey (1984) for orthopyroxene and clinopyroxene and from McKenzie and O'Nions (1991) for olivine, spinel, and garnet. Normalizing values are from Sun and McDonough (1989).

Fig. 15. Petrogenetic model for post-collisional ultrapotassic magmatism and incorporation of related xenoliths in the Eslamieh Peninsula. Slab rollback could have induced mantle uprising and associated melting of an enriched peridotite mantle source, which produced high-MgO and low-MgO ultrapotassic magmatic rocks, respectively, in the Eslamieh Peninsula