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

Fatemeh Sepidbar, Seyed Masoud Homam, Qiao Shu, Richard M. Palin, Pouya Besharati, Michele Lustrino, *Seyed Masoud Homam

ARTICLE INFO

Keywords:
- Ultrapotassic magmatic rocks
- Mantle peridotite
- Xenoliths
- Eslamieh Peninsula
- NW Iran

ABSTRACT

The Eslamieh Peninsula in NW Iran exposes Miocene potassic to ultrapotassic volcanic rocks that are often associated with a variety of ultramafic xenoliths. Based on geochemical features, the rocks investigated in this study can be subdivided into two main groups: 1) MgO-rich (8.0–13.9 wt%) basic (SiO$_2$ = 46.1–50.9 wt%) rocks, which comprise mafic lamprophyres and pyroxenite xenoliths, and 2) MgO-poor (1.5–6.4 wt%) basic to intermediate (SiO$_2$ = 48.2–61.2 wt%) rocks, 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 LREE over HREE (e.g., La/Yb = 17–55), 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 peaks in primitive mantle-normalised diagrams (e.g., Nb/Nb* mostly <0.7). Both high- and poor-MgO samples have strong radiogenic (87Sr/86Sr), ratios (0.7078–0.7086), below the Chondritic Uniform Reservoir (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. The major oxide and trace element contents the high-MgO and high-CaO rocks indicate derivation from a subcontinental lithospheric mantle source, variably metasomatised by fluids and melts released by a subducted slab in a mantle wedge. Primitive melts evolved via mafic mineral fractionation and possible assimilation of lower crust, generating the SiO$_2$ higher and CaO-MgO-poorer 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.

1. 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 (Fig. 1A). 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; Sepidbar et al., 2021; Rabiee et al., 2022). Magma underplating and peridotite-melt reactions (Khedr et al., 2014; Moghadam et al., 2014) have been proposed as two of the 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

* Corresponding authors.
E-mail addresses: f.sepidbar@ut.ac.ir (F. Sepidbar), homam@um.ac.ir (S.M. Homam).

https://doi.org/10.1016/j.lithos.2023.107359

Received 23 March 2023; Received in revised form 12 September 2023; Accepted 13 September 2023
Available online 15 September 2023
0024-4937/Crown Copyright © 2023 Published by Elsevier B.V. All rights reserved.
the focus of much interest in recent years owing to their importance for understanding the nature of the upper mantle (Moghadam et al., 2014; Pang et al., 2013; Rabiee et al., 2022; Sepidbar et al., 2021). 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. Aghazadeh et al., 2015; Chung et al., 2009; Guo et al., 2015; Lustrino et al., 2019, 2022; Zhao et al., 2009). 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 (Zhao et al., 2009); and (3) rollback and break-off of a subducted slab of the continental lithosphere (Chen et al., 2015; Guo et al., 2015; Moghadam et al., 2018; 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 K-rich magmas in a back-arc setting.

Fig. 1. (A) Simplified map showing Turkish-Iranian High Plateau, evidencing in yellow the areas where Cenozoic igneous rocks crop out. (B) Simplified geological map of Iran showing Cenozoic magmatic rocks and the position of the Eslamieh Peninsula. (C) Geological map of the Eslamieh Peninsula. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Post-collisional potassic to ultrapotassic igneous rocks in the Eslamieh (known also as Eslami or Esłamy Peninsula or as the “Saray volcano”; e.g., Moghadam et al., 2014; Aghazadeh et al., 2015), 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 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; Forster et al., 2019; 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 (Aghazadeh et al., 2015; 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., Chiu et al., 2013; Dilek et al., 2010). 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, 2013b).

In the eastern part of the Turkish-Iranian High Plateau, widespread Late Neoproterozoic-Early Cambrian (Gadamian) rocks crop out. Their large variations in initial 87Sr/86Sr ratios and εNd 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 Soltanian, 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 saline-clay deposits of the Urumieh Lake (Moghadam et al., 2014; Moine-Vaziri et al., 1991). Late Miocene potassic and ultrapotassic magmatic rocks, containing a variable cargo of mantle and crustal xenoliths, occur in the eastern parts of the Turkish-Iranian High Plateau (Khezerlou et al., 2017; Moine-Vaziri et al., 1991), along the Urumieh-Dokhtar Magmatic Arc (UDMA), and in the Eslamieh Peninsula (Moayyed et al., 2008; 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, 2013b; Fedele et al., 2022, 2023; Lustrino et al., 2021; 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 (Aghazadeh et al., 2015; Khezerlou et al., 2017; Fig. 1A, B). The Eslamieh Peninsula magmatic rocks were mostly emplaced at 11–10 Ma (Aghazadeh et al., 2015; Pang et al., 2013), are characterized by alkaline to strongly alkaline compositions (essentially potassic to ultrapotassic), 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 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 saline-clay deposits of Lake Urumieh (Fig. 1C).

2.1.1. Volcanic rocks

The volcanic rocks of Eslamieh Peninsula outcrop over an area of ~150 km² and have been subdivided into six major lithologies: (1) pyroxenite-glimmerite, nepheline-syenite and monzonite xenoliths hosted by the volcanic rocks; (2) potassic 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 and 5 cm to even >1 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), contain plenty of pyroxenite and glimmerite xenoliths (Figs. 2-3). Trachytes are intruded in the Early Miocene pyroclastic rocks and overlie with Late Miocene conglomerate type sediments with pebbles of older rock units and lahar deposits (Fig. 2A), and contain pyroxenite xenoliths (Fig. 2B) They comprise of ~10–15 vol% phenocrysts of euhedral sanidine and light to dark green clinopyroxene. Phonolites are juxtaposed with trachytic and pyroclastic rocks to the south and north, respectively (Fig. 1C) and contain plenty of pyroxenite xenoliths (Fig. 2C). 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. 2D), 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.

2.1.2. Ultramafic xenoliths

The present study reports an investigation on pyroxenites and glimmerites, found as xenoliths in trachytes (Fig. 2B), phonolite lavas (Fig. 2C) and felsic lamprophyric dikes, respectively. They occur with an irregular shape with a size ranging from <1 cm up to 10 cm, and show sharp contacts with the host magmas.
3. Analytical techniques

Major oxide 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 clinoxyroxene 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 clinoxyroxene 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. BHVO-2 glass was measured as an unknown to verify data accuracy. The analytical uncertainties are better than 5% for most trace elements. Detailed analytical procedures have been reported by Liu (2012).

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).

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 7700× 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 corrections were applied. Detailed analytical procedures have been reported by Liu (2012).
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 91,500 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 91,500 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/).

4. Analytical results

4.1. Petrography

The investigated samples from the Eslamieh Peninsula were classified into trachyte and phonolite lavas, mafic and felsic lamprophyric dikes, and xenoliths based on petrographic considerations. 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. 3A), 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 5 cm), as well as forming in the groundmass, where it is aligned to form a trachytic texture (Fig. 3B). 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 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%), together with microlites of sanidine (~5–10 vol %), in a plagioclase-free apatite-bearing groundmass (Fig. 3C-D). These characteristics are typical of minettes (i.e., phlogopite-rich, plagioclase-poor/free lamprophyres), as already identified by Aghazadeh et al. (2015). Phlogopite is the major phenocryst in mafic lamprophyres, with
<table>
<thead>
<tr>
<th>Sample</th>
<th>Age (Ma)</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>P₂O₅</th>
<th>LOI</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>024-H</td>
<td>10.72</td>
<td>51.9</td>
<td>13.4</td>
<td>12.6</td>
<td>0.17</td>
<td>4.57</td>
<td>4.98</td>
<td>3.80</td>
<td>7.69</td>
<td>1.30</td>
<td>5.25</td>
<td>101</td>
</tr>
<tr>
<td>026-H</td>
<td></td>
<td>47.3</td>
<td>16.2</td>
<td>12.7</td>
<td>0.38</td>
<td>6.34</td>
<td>4.88</td>
<td>2.48</td>
<td>8.16</td>
<td>1.61</td>
<td>6.58</td>
<td>101</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sample</th>
<th>Hf</th>
<th>Ta</th>
<th>Sc</th>
<th>La</th>
<th>Pr</th>
<th>Nd</th>
<th>Eu</th>
<th>Dy</th>
<th>Ho</th>
<th>Er</th>
<th>Yb</th>
<th>Lu</th>
<th>Prn</th>
</tr>
</thead>
<tbody>
<tr>
<td>024-H</td>
<td>20.3</td>
<td>4.84</td>
<td>22.3</td>
<td>39.6</td>
<td>10.7</td>
<td>42.1</td>
<td>1.93</td>
<td>5.14</td>
<td>0.57</td>
<td>2.46</td>
<td>3.05</td>
<td>0.32</td>
<td>0.708239</td>
</tr>
<tr>
<td>026-H</td>
<td>21.2</td>
<td>4.24</td>
<td>22.3</td>
<td>39.6</td>
<td>10.7</td>
<td>42.1</td>
<td>1.93</td>
<td>5.14</td>
<td>0.57</td>
<td>2.46</td>
<td>3.05</td>
<td>0.32</td>
<td>0.708413</td>
</tr>
</tbody>
</table>

(continued on next page)
Table 1 (continued)

| Series Name | 024–I(H) 026(H) | 20(H) | 19(H) | 18(H) | 16(H) | 17(H) | 15(H) | 14(H) | 13(H) | 12(H) | 11(H) | 10(H) | 9(H) | 8(H) | 7(H) | 6(H) | 5(H) | 4(H) | 3(H) | 2(H) | 1(H) | 1(X) | 2(X) | 3(X) | 4(X) |
|-------------|-----------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| Rock Type   | Phonolite       | Trachyte | Felsic lamprophyre | Mafic lamprophyre | Pyroxenite | Xenolith |
| U           | 1.2             | 1.1   | 27    | 14    | 22    | 22    | 7     | 12    | 16    | 21    | 27    | 13    | 15    | 14    | 18    | 13    | 19    | 12    | 13    | 6    | 8    | 19    |
| Ga          | 21.6            | 23.5  | 19.0  | 18.0  | 22.0  | 21.0  | 17.0  | 20.0  | 20.0  | 17.0  | 19.0  | 16.0  | 17.0  | 15.0  | 18.0  | 19.0  | 13.0  | 14.0  | 12.0  | 15.0  | 13.0  | 11.1  | 14.4  | 12.9  | 5.26  |
| Nb/La       | 2.3             | 2.4   | 0.7   | 0.9   | 0.8   | 0.7   | 0.4   | 0.6   | 0.9   | 0.8   | 0.9   | 0.7   | 0.7   | 0.7   | 0.8   | 0.7   | 0.6   | 0.8   | 0.6   | 0.4   | 0.5   | 0.5   | 0.5   | 0.8   | 0.1   | 0.1   |
| Nb/Ta       | 19.1            | 21.5  | 75.0  | 16.3  | 17.8  | 22.3  | 11.3  | 16.7  | 21.7  | 20.3  | 29.5  | 19.5  | 25.5  | 18.3  | 18.3  | 19.5  | 23.5  | 30.5  | 17.3  | 31.0  | 21.0  | 15.3  | 17.2  | 20.2  | 14.6  | 11.4  |
| Nb/Yb       | 40.3            | 55.2  | 37.5  | 32.5  | 29.7  | 33.5  | 17.0  | 25.0  | 32.5  | 30.5  | 29.5  | 19.5  | 25.5  | 27.5  | 27.5  | 49.0  | 23.5  | 30.5  | 26.0  | 15.5  | 21.0  | 23.9  | 14.3  | 17.9  | 2.7   | 2.7   |
| Nb/Rb       | 0.9             | 0.5   | 0.2   | 0.4   | 0.4   | 0.2   | 0.1   | 0.2   | 0.2   | 0.3   | 0.3   | 0.2   | 0.2   | 0.2   | 0.2   | 0.2   | 0.2   | 0.5   | 0.3   | 0.2   | 0.2   | 0.3   | 0.3   | 0.3   | 0.3   | 0.1   | 0.1   |
| Nb/Th       | 2.8             | 7.3   | 1.2   | 1.7   | 1.2   | 1.3   | 0.7   | 0.6   | 1.5   | 0.9   | 1.1   | 0.6   | 1.3   | 1.3   | 1.3   | 1.0   | 1.0   | 1.4   | 1.2   | 1.1   | 1.2   | 1.2   | 1.6   | 4.7   | 1.7   | 2.7   |
| La/Lu       | 121.5           | 169.0 | 212.6 | 161.7 | 178.8 | 208.7 |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       | 2.7   | 2.7   |
| La/Sr       | 4.5             | 5.0   | 7.3   | 6.0   | 7.3   | 7.5   | 6.4   | 6.5   | 6.9   | 7.2   | 7.0   | 6.6   | 7.7   | 6.2   | 8.9   | 5.0   | 5.9   | 4.2   | 4.8   | 5.6   | 5.1   | 4.5   | 4.7   | 4.7   | 3.6   | 5.7   |
| La/Yb       | 17.3            | 23.2  | 55.0  | 36.0  | 38.7  | 48.5  | 41.5  | 39.0  | 38.0  | 36.0  | 31.5  | 29.5  | 34.5  | 37.0  | 35.5  | 75.0  | 38.5  | 39.5  | 41.0  | 36.5  | 45.5  | 44.5  | 29.6  | 23.6  | 24.9  | 27.6  |
| Sm/Yb       | 240.2           | 198.8 | 617.0 | 878.0 | 522.7 | 806.5 | 744.5 | 1211.5| 862.0 | 786.0 | 828.5 | 934.5 | 711.0 | 757.0 | 629.0 | 1625.0| 562.5 | 662.5 | 577.0 | 581.5 | 618.0 | 572.5 | 312.0 | 487.5 | 230.9 | 136.7 |
| Sr/Y         | 13.1            | 8.1   | 15.0  | 29.8  | 17.2  | 25.2  | 19.3  | 33.7  | 26.1  | 28.6  | 33.1  | 38.1  | 27.9  | 24.0  | 22.9  | 26.2  | 17.3  | 14.9 | 14.1 | 17.4 | 16.5 | 16.6 | 10.0 | 19.8 | 7.4   | 4.8   |
| Eu/Eu*      | 0.77            | 0.84  | 0.47  | 0.62  | 0.44  | 0.53  | 0.80  | 0.58  | 0.61  | 0.68  | 0.36  | 0.67  | 0.72  | 0.58  | 0.72  | 0.83  | 0.89  | 0.42  | 0.64  | 0.76  | 0.62  | 0.59  | 0.72  | 0.80  | 0.75  | 0.91  |
| ΔNb         | -0.7            | -0.9  | -0.4  | -0.1  | -0.6  | -0.3  | -0.6  | -0.4  | -0.4  | -0.6  | -0.6  | -0.5  | -0.3  | -0.3  | -0.5  | -0.6  | -0.5  | 0.0   | -0.3  | -0.2  | 0.2   | -0.2  | -0.3  | 0.0   | 0.0   |

87Sr/86Sr: Obtained Initial 87Sr/86Sr based on age of 10.72 Ma.
143Nd/144Nd: Obtained Initial 143Nd/144Nd based on age of 10.72 Ma.
176Hf/177Hf: Obtained Initial 176Hf/177Hf based on age of 10.72 Ma.
ASI: (alumina saturation index): molar Al2O3/(CaO + Na2O + K2O).
La N to Lu N: Rare earth elements values in chondrite normalised based on Sun and McDonough (1989).
Mg#: (100 × Mg/(Mg + Fe2+)).
ΔNb; 1.74 + log (Nb/Y) - 1.92log(Zr/Y).
Eu/Eu*, is calculated as: EuCN/(SmCN*GdCN)0.5.
sizes up to 1 cm, followed by clinopyroxene, magnetite andapatite. 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. They carry abundant irregularly-shaped pyroxenites and glimmerite xenoliths up to 25–30 cm in size.

4.1. 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), 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 lamprophyres mainly consist of phlogopite (~50–80 vol%) and clinopyroxene (~20–40 vol%) (Fig. 3F). 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.

4.2. 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. 4a; Supplementary Appendix S1) and 10.64 ± 0.07 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 40Ar/39Ar range reported in the literature (10.33–10.93 Ma; Pang et al., 2013; Moghadam et al., 2014; Aghazadeh et al., 2015; Supplementary Appendix S1).

4.3. Mineral compositions

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

Clinopyroxenes are 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 (Wo46-49Fs40-42Fs3-7.1) and having Mg# (Mg# = Mg/(Mg + Fe2+)) 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 (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 (Wo44.69En40.48Fs11.1), with minor acmite (<2.5%) and kushiroite (Al2O3 = 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 Wo45.71En43.94Fs8.34 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 Kushiro 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) roughly similar to that of the xenoliths (40–129 ppm; Supplementary Appendix S2a). All pyroxenes show similar LREE enrichment over HREE (LaN/YbN = 3.1–5.7 and 3.3–8.1 for trachyte and phonolite, respectively), with slight enrichment of MREE over HREE (LaN/YbN = 9.7 and 11.6 for trachyte and phonolite, respectively), with slight enrichment of MREE over HREE (LaN/YbN = 9.7 and 11.6 for trachyte and phonolite, respectively). However, higher MgO contents mostly occurred in clinopyroxene from pyroxenite xenoliths (Supplementary Appendix S2a), whereas those from pyroxenite xenoliths (Supplementary Appendix S2b) have lower trace element contents (e.g., average ∑REE = ~98 ppm) than low-MgO zones (average ∑REE = 185 ppm). Clinopyroxenes have a clear depletion in highly incompatible trace elements (e.g., average Rb

---

**Fig. 4.** (A) Inverse and (B) weighted mean 206Pb/238U age plots for zircons from trachytic samples from Eslamieh Peninsula. Details are reported in Supplementary Appendix 1.
Lithos 458–459 (2023) 107359

Phlogopite was analysed from trachytic lavas (Supplementary Appendix S2c) as well as pyroxenite and glimmerite xenoliths (Supplementary Appendix S2d). Phlogopite in trachytic lavas have TiO$_2$ in the 1.7–3.7 wt% range and MgO-rich compositions (MgO = 18.8–25.1 wt%; Mg#$ = 0.82–0.91$). Compositions are similar to phlogopite in peridotite xenoliths hosted in alkali-rich magmas worldwide (Figs. 7E; Lustrino et al., 1999; Pearson et al., 2003). The phlogopite found in pyroxenite xenoliths show lower MgO (15.2–17.9 wt%) and Mg# (0.68–0.76), coupled with higher TiO$_2$ (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 4585–41,771 and 5466–48,405 ppm), respectively (Supplementary Appendix S2b).

Fig. 5. BSE-SEM images of clinopyroxene in a trachyte (A-1), phonolite (B-1), and a pyroxenite xenolith (C-1); (A-2) Rim-to-core MgO variation of clinopyroxene in a trachyte; (B-2) in a phonolite, and in a pyroxenite xenolith.
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 S4. The investigated rocks plot in the ultrapotassic field in the K$_2$O vs. Na$_2$O diagram, being characterized mostly by K$_2$O > 3 wt% and K$_2$O/Na$_2$O > 2 (Fig. 8A; Supplementary Appendix 4). 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 (K$_2$O + Na$_2$O wt%) vs. SiO$_2$ 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

Fig. 6. CI chondrite-normalised (after King et al., 2020) REE and primitive mantle-normalised (after Lyubetskaya and Korenaga, 2007) incompatible element diagrams for clinopyroxene in trachytes (A-B), phonolites (C–D), and pyroxenite xenoliths (E-F).
phonolite field. On a Zr/TiO$_2$ vs. Nb/Y diagram (Pearce, 1996), most of the samples cluster in a relatively restricted area between the trachyte-trachyandesite fields (Fig. 8 C).

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 SiO$_2$ (46.0–50.7 wt%) and generally lower Na$_2$O (excluding one xenolith) contents, coupled with higher Mg# [0.65–0.94; Mg# = Mg/(Mg + Fe$^{2+}$), assuming Fe$_{2}$O$_3$/(FeO + Fe$_{2}$O$_3$) = 0.15], hereafter introduce as high-MgO subgroup compared with the trachyte, phonolite and felsic lamprophyre, known as low-MgO subgroup (Supplementary Appendix S4).

The new rock compositions reported in this study almost completely overlap the literature compositions of Eslamieh samples. They show either roughly positive (TiO$_2$, Fe$_{2}$O$_3$tot, CaO, and P$_2$O$_5$) or negative correlations (SiO$_2$, Al$_2$O$_3$, and alkalis) with MgO (Fig. 9; Supplementary Appendix S4).

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$_2$O$_3$ vs. Mg# diagrams for clinopyroxene in volcanic host rocks and xenoliths; (E) 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). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Appendix S4). 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-Ta-Ti, coupled with positive peaks at Pb and K, and a general enrichment of LILE compared to HFSE (Fig. 10; Supplementary Appendix 4). 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-normalised REE patterns for mafic and felsic lamprophyres (Fig. 10; Supplementary Appendix 4), with the former showing slightly higher absolute concentrations than the latter. Overall, a large inter-elemental fractionation overlap is observed, with La/Nb/Yb ranging from 25 to 31 and from 23 to 51 for mafic and felsic lamprophyres, respectively. Both lamprophyre types are characterized by strong negative Eu anomalies (Eu/Eu* = 0.42–0.76 and 0.36–0.80 for the mafic and felsic variants, respectively). The trachytic rock group shows complete compositional overlap with the lamprophyres (LaB/YbB = 20–37; Eu/Eu* = 0.36–0.80; Fig. 10), whereas the pyroxenite xenoliths (LaB/YbB = 17–20; Eu/Eu* = 0.72–0.91) and the phonolites (LaB/YbB = 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).

4.5. 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 and in Supplementary Appendix 4. Eslamieh lavas have strongly radiogenic ([87Sr/86Sr]i ratios (0.70824–0.70863) and strongly unradiogenic Nd ([143Nd/144Nd])i ratios (0.51235–0.51240). 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., 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 characterized by less radiogenic ([87Sr/86Sr])i = 0.70780, and more radiogenic ([143Nd/144Nd])i = 0.51245. The initial εHf 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).

5. Discussion

The origin and evolution of magmas emplaced along active margins comprise a series of processes that include (e.g., Eyuboglu et al., 2013a, 2013b; 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, 1997b; Khedr et al., 2010; Khedr and Arai, 2009); (2) variable partial melting of the metasomatized mantle (e.g., Pearce and
Fig. 9. Selected major oxide (wt%) vs. MgO (wt%) contents illustrating the compositional range of the magmatic rocks in the Eslamieh Peninsula. All the major element data were recalculated to 100 wt% on a volatile-free basis. Data sources and symbols are as in Fig. 8.
Fig. 10. CI chondrite-normalised REE-normalised (King et al., 2020) and primitive mantle-normalised (Lyubetskaya and Korenaga, 2007) incompatible element diagrams for the Eslamieh ultrapotassic magmatic rocks: (A-B) phonolites, (C–D) trachytes, (E-H) felsic lamprophyres, (G-H) mafic lamprophyres, (I-J) ultramafic xenoliths.
5.1. Origin of the pyroxenite xenoliths

Pyroxenites comprise a volumetrically minor portion of the upper mantle, although their presence can significantly influence the composition of mantle-derived partial melts. These compositions range from orthopyroxenites through websterites and 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 (1) high-pressure cumulates from mantle-derived melts (Xu, 2002); (2) gabbroic cumulates from mantle-derived melts, which were transformed to pyroxenites during subsequent subduction (Yu et al., 2010), or (3) melt and/or fluid-peridotite reactions (Hu et al., 2016). Pyroxenites usually preserve a more extensive reaction history compared to the possibly associated mantle-derived melts, which were transformed to pyroxenites during subsequent subduction (Yu et al., 2010), or (3) melt and/or fluid-peridotite reactions (Hu et al., 2016). Pyroxenites in this study are characterized by relatively low SiO$_2$ (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, pyroxenites from pyroxenite xenoliths from Eslamieh Peninsula have low Al$_2$O$_3$ contents (3.6–5.6 wt% wt%), and their CI chondrite-normalised REE patterns are characterized by small 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 are characterized by high Al$_2$O$_3$ (>12 wt%) and positive Eu anomalies (Yu et al., 2010). The $\delta^{87}$Sr/$\delta^{86}$Sr higher than Bulk Silicate Earth (BSE) estimate, coupled with $^{143}$Nd/$^{144}$Nd lower than CHUR concord in excluding a MOR-related protolith for these compositions. Clinopyroxenites 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; ).
Hawaiian data (Sobolev et al., 2005). separating peridotite- and pyroxenite-source primary magmas are after Herzberg (2011). (B) Plot of Ni vs. MgO. Broken black lines pyroxenite and peridotite partial melts in (A) and (B) are outlined based on Herzberg and Asimow (2008) and Guo et al. (2015). The compositional fields of rocks in the Eslamieh Peninsula. The CaO contents for peridotite partial melts Fig. 12. F. Sepidbar et al.

dehydration melting at ~1 GPa, 1050 ºC (Conceição and Green, 2004). The Sr/Pb ages of mica: ~10.7 ± 0.02 Ma; Agha-Panah et al., 2009; Ersoy et al., 2008). Field observations (e.g., Moayyed et al., 2008; Moghadam et al., 2014; this study) coupled with radiometric age constraints, indicate that phonolites (U–Pb ages of zircon: ~10.7 ± 0.2 Ma; 11.0 ± 0.1 Ma; Pang et al., 2013) are as same with the lamprophyres (Ar–Ar ages of mica: ~10.98 ± 0.4 Ma and 10.35 ± 0.02 Ma; Aghazadeh et al., 2015) and 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).

The co-occurrence of mafic and felsic lamprophyres is an unusual feature of the Eslamieh Peninsula, although such a similar pairing was
also reported in Buell Park, Arizona by Roden and Smith (1971). These authors suggested 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 a similar source. 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$_2$O$_3$, but lower MgO, CaO, TiO$_2$, P$_2$O$_5$, Fe$_{2}$O$_{3}$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 in the trachytes 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

Fig. 13. (A-B) Th/Yb vs. Nb/Yb; (C) Th/Nd vs. Ba/La; (D) Th/La vs. Sm/La; and (E) whole-rock Rb/Sr vs. Ba/Rb. The arrows labelled with phlogopite and amphibole in (ED) refer to the effect of increasing presence of the phases in the mantle source region. The orange field in (D) shows the compositional variation of arc magmatic rocks related to slab-derived mantle metasomatism (Plank et al., 2007); the blue field covers the compositional trend of Tethyan-realm lamproites (Prelević et al., 2013; Tommasini et al., 2011). 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 Fig. 8. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Chapman and Powell (1976), who considered K-feldspar megacrysts in lamprophyres to be xenocrystic in origin.

6. Geodynamic implications

Potassic and ultrapotassic magmas 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 (Mohgadam and Stern, 2011), implying that the late Cenozoic magmatism in the eastern parts of the Turkish-Iranian High Plateau and northwestern part of the UDMA occurred in a post-collisional setting. Consequently, the most likely tectonic setting is supra-subduction sub-continental lithospheric mantle wedge that experienced small degree partial melting (e.g., Moghadam et al., 2014; Aghazadeh et al., 2015). The presence of high- and low-MgO pyroxene with different REE distribution (Fig. 6) within the abundant pyroxenite xenoliths in the Eslamieh Peninsula confirm the process of mantle refertilization by melt/fluid metasomatism (e.g., Melchiorre et al., 2019).

Based on whole-rock compositions (including Sr-Nd-Hf isotopic ratios), we propose that the ultrapotassic magmas of the Eslamieh Peninsula mainly originated from partial melting of a metasomatized peridotitic source. Our conclusion agrees with that of Moghadam et al. (2014), who proposed that Eslamieh rocks derived from mantle sources modified by fluids Ca-rich plagioclase in the investigated rocks. The negative Eu/Eu* ratios would simply indicate derivation from mantle sources of Eslamieh ultrapotassic magmas, considering the absence of crustal lithologies in the Eslamieh Peninsula lavas. Underplating of the subducted slab in the mantle wedge, giving rise to production of high-MgO ultrapotassic magmas. Low-degree partial melting of the metasomatized mantle source, infiltrated heterogeneously with melt or fluid released by previous subducted slab in the mantle wedge, 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 and Jackson, 1995). 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).

7. 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 xenoliths. The xenoliths investigated in the present study are pyroxenites and gilmerites.

(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.

**Declaration of Competing Interest**

We declare that we have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

**Acknowledgements**

This work was supported by the Ferdowsi University of Mashhad and Institute of Geochemistry, Chinese Academy of Sciences (China), for which we are thankful. We thank Prof. Greg Shellnutt for editorial handling and Prof. Mohamed Zaki Khedr and one anonymous reviewer for their positive comments and suggestions, allowing us to greatly
improve the manuscript. The first author acknowledges support from Ferworsh University of Mashhad for assistance during field work. ML acknowledges funds PRIN2017 Project 20177BBX42.005 and Fondi Sapienza Ateneo 2021 and 2022.

Appendix A Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.lithos.2023.107359.

References


