Late Miocene K-rich volcanism in the Eslamieh Peninsula (Saray), NW Iran: Implications for geodynamic evolution of the Turkish–Iranian High Plateau

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A B S T R A C T

Post-collisional volcanism in northwestern Iran is represented by the Saray high-K rocks including leucite-bearing under-saturated and leucite-free silica saturated rocks. We report Ar–Ar age data which constrain the age as ca. 11 Ma (Late Miocene). Most of clinopyroxene phenocrysts from the volcanic rocks have complex oscillatory zoning, with high Ti and Al cores, low Ti and high Al mantled clinopyroxenes, grading into low Ti and Al outer rims. All the rocks are highly enriched in incompatible trace elements and have identical Sr–Nd–Pb isotopes. Enrichment in incompatible elements and other geochemical features for the Saray lavas suggest a metasomatized sub-continental lithospheric mantle (SCLM) as the magma source. The negative Nb–Ta–Ti anomalies for the Saray lavas compare with the features of subduction-related magmatism with negligible contamination with ancient crustal components. The highly radiogenic \(^{87}\)Sr/\(^{86}\)Sr and \(^{207}\)Pb/\(^{204}\)Pb isotopic values of the Saray lavas imply the involvement of slab tectonics sediments and/or a continental lithosphere. Isotopically, the volcanic rocks define a binary trend, representing 5–8% mixing between the primary mantle and sediment melts. Our melting models suggest residual garnet in the source and are incompatible with partial melting of amphibole and/or phlogopite-bearing lherzolites, although the complex geochemical features might indicate the result of mixing between melts produced by different sources or a homogenous melt passing through a compositionally-zoned mantle during multiple stages of partial melting and melt migration. The geochronological, geochemical and isotopic data for the Saray rocks suggest that these Late Miocene magmas were derived from a small degree of partial melting of subduction-metasomatized (subcontinental) lithospheric mantle source in a post-collisional setting.

Keywords: High-K lavas Geochemistry, Sr–Nd–Pb isotopes, Post-collisional magmatism Iran

1. Introduction

The northward migration of the Afro-Arabian plate caused subduction during the late Cretaceous to Miocene beneath the Iranian block, followed by collision between the Anatolian–Iranian and Arabian plates during middle to late Miocene (Berberian and King, 1981; Alici Sen et al., 2004; Chiu et al., 2013). This continental collision was accommodated by crustal shortening and thickening in eastern Anatolia and northwestern Iran (Sengor and Yilmaz, 1981; Temel et al., 1998; Alici Sen et al., 2004; Dilek et al., 2010). The northwestern Iran–eastern Mediterranean region (Turkish–Iranian High Plateau) is a tectonically active plateau between the converging Arabia and Eurasia. Cenozoic magmatic rocks occur extensively in the Turkish–Iranian High Plateau, northeast of the Bitlis–Zagros suture zone (Dilek et al., 2010) (Fig. 1). The temporal distribution of the magmatism in this region shows different phases of magma generation with distinct geochemical signatures mainly in the late Eocene, late Miocene and Plio–Quaternary (Dilek et al., 2010; Eyuboglu et al., 2010, 2011a,b,c,d,e, 2012, 2013a,b). Mantle-derived volcanic rocks with oceanic-island basalt (OIB)-like geochemical features with and/or without subduction fingerprints are commonly found in the Turkish–Iranian High Plateau (e.g., Aldanmaz et al., 2006; Dilek and Altunkaynak, 2009; Kheirkhah et al., 2009; Elitok et al., 2010; Allen et al., 2013).
The late Miocene magmatic rocks with calc-alkaline and shoshonitic (or ultrapotassic) characteristics are found mainly along the UDMA (Chiu et al., 2013) and occur in the northern parts of the Turkish–Iranian High Plateau (Moine-Vaziri et al., 1991; Dilek et al., 2010; Eyuboglu et al., 2012). Plio-Quaternary magmatic rocks are represented by alkaline rocks with or without subduction overprints and by shoshonitic rocks in several regions of the UDMA, but mostly in NW Iran, in the Turkish-Iranian High Plateau (e.g., Keskin, 2003; Kheirkhah et al., 2009; Elitok et al., 2010; Eyuboglu et al., 2012; Allen et al., 2013).

The study area is situated between the Bitlis–Zagros Zone and eastern Pontides–Lesser Caucasus–Alborz magmatic arc. Although there is a consensus on a northward subduction model for the geodynamic evolution of the Bitlis–Zagros Zone in the southern part of the study area, the evolution of the eastern Pontides–Lesser Caucasus–Alborz magmatic arc in the north is still controversial due to lack of systematic geological, geochronological and geochemical data. The most popular idea is the magmatic arc developed by the northward subduction of the Paleotethys oceanic lithosphere, which was situated to the south of the arc, until the Eocene (e.g., Adamia et al., 1977; Okay and Sahintürk, 1997; Dilek et al., 2010). An earlier interpretation by Sengor and Yilmaz (1981) proposed that Paleotethys was located to the north of the magmatic arc, and hence southward subduction operated from the Paleozoic until the Dogger, and that northward subduction occurred subsequently from the late Mesozoic until the Eocene. The third model includes those of Dewey et al. (1973), Bektas et al. (1999), Chorowicz et al. (1998), Eyuboglu et al. (2006, 2007, 2010) and Eyuboglu (2010) who favor a southward subduction of the Tethys oceanic lithosphere, which was situated to the north of magmatic arc, continued uninterruptedly from the Paleozoic till the late Cenozoic.

Adakitic volcanism is minor in the region north of the Bitlis–Zagros suture zone in the northern and southern parts (e.g., Jahangiri, 2007; Omrani et al., 2008; Chiu et al., 2013). However, different pulses of magmatism are widespread in the Eastern Pontides–Lesser Caucasus and eastern Iranian–Azerbaijan regions, which have been correlated with the northward subduction of the southern Neotethyan Ocean (Yilmaz et al., 1997). These rocks vary in composition from calc-alkaline to alkaline/adakitic characteristics with temporal variations from Paleocene to Plio-Quaternary. The occurrence of magmatic rocks in the Eastern Pontides–Lesser Caucasus and eastern Iranian–Azerbaijan regions have posed important questions related to the geodynamic evolution of these regions. Recent comprehensive geological, geochemical, and geochronological studies (e.g., Dokuz, 2011; Eyuboglu et al., 2011a,b,c,d,e, 2012, 2013a, 2013b; Maden, 2013) on the Cenozoic magmatism revealed that the adakitic magmatism occurred in two different cycles in the eastern Pontides–Lesser Caucasus–Alborz magmatic arc. The first phase was during the late Paleocene–early Eocene and the second one during late Miocene. The origin of the late Paleocene–early Eocene adakitic rocks, which are well exposed in the southern part of the eastern Pontides orogenic belt, has been correlated to slab window processes in a southward subduction zone (Eyuboglu et al., 2011a,b,c,e, 2013a,b). The late Miocene adakitic rocks, which are well exposed in the eastern part of the eastern Pontides orogenic belt and also Caucasus, were generated by partial melting of the mafic lower crust after the collision between Lesser Caucasus and Greater Caucasus (Eyuboglu et al., 2012). These studies suggest that the slab break-off of the southward subducted oceanic lithosphere and inflow of hot asthenosphere into the mantle wedge caused partial melting of subduction-modified mantle wedge and led to the formation of Pliocene and Pleistocene alkaline magmatic rocks exposed along the northern edge of the eastern Pontides–Lesser Caucasus–Alborz magmatic arc (Eyuboglu et al., 2018). Although several studies have highlighted the temporal and spatial aspects of magmatism in the Turkish–Iranian High Plateau, showing different pulses of magma generation mainly in the late Eocene, late Miocene and Plio–Quaternary and with distinct geochemical signatures, NW Iran has not been investigated in detail. A recent report shows that Late Eocene and Oligocene (ca. 38 to 23 Ma) shoshonitic to adakitic plutons (with an age progression from SE to NW) are common in NW–N Iran (Castro et al., 2013). It is important to trace the temporal distribution of the various pulses of magmatism and the triggering mechanisms in NW Iran, in order to evaluate the post-collisional magmatism in this region.

In this paper we present: 1) a detailed classification of the high-K volcanic rocks in the Eslamieh Peninsula (Saray) based on mineral and whole rock geochemical studies; 2) Ar–Ar dating of the rocks and 3) Sr–Nd–Pb isotope geochemistry. Based on the results, we evaluate the petrogenesis of the high-K rocks (leucite-free or leucite-bearing) and its implications on post-collisional magmatism in the late Miocene.

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We then interpret our data in the geodynamic framework of the Turkish–Iranian High Plateau in order to constrain the mantle source and magma evolution of high-K volcanic rocks.

2. Geological setting and field relationships of high-K volcanic rocks

The late Miocene (Tortonian) high K-volcanic rocks are exposed mainly in the Eslamieh Peninsula (Saray), to the southeast of Urumieh city (Figs. 1 and 2). The contact relationships with older strata, including the magmatic and metamorphic basement (such as the Gushchi A-type granites and Cadomian granitic gneisses), are mostly covered by younger salty-clay deposits of the Urumieh Lake. The late Miocene high-K Saray volcanic rocks appear to crosscut the Iranian basement. The Cadomian granitic gneisses show similarity to those described in central Iran by Hassanzadeh et al. (2008). The Gushchi A-type granites are associated with OIB-

![Fig. 2. Simplified geological map of the Eslamieh (Saray) Peninsula. Modified after Moine-Vaziri et al. (1991).](image-url)
type alkaline gabbros and show intrusive contacts with these alkaline rocks. The gabbroic rocks yielded Ar–Ar ages of ca. 316 Ma, although they seem to be older than 360 Ma, indicating a phase of Ar resetting (Shahabi, 2013).

Field observations show that the volcanic activity in the Saray can be divided into six major stages as follows. (i) An older, probably early Miocene (Aquitanian–Langhian) plutonism represented by nepheline–syenite and monzonite xenoliths within the high-K volcanic rocks. (ii) Late Miocene (Tortonian) explosive activity initiated with a major caldera-forming episode. This magmatic episode is composed of basanite, leucite–tephrite and pyroclastic rocks in the lower volcanic unit and phonolite, trachyte and analcime-bearing basanite in the upper volcanic unit (Moine-Vaziri et al., 1991). (iii) A phase of highly explosive activity with deposition of a thick pyroclastic sequence. The rock fragments in these pyroclastic assemblages are similar to the volcanic rocks of stage (ii) or are petrographically (see Section 3) different. Clinopyroxene–phlogopite–feldspar-rich cumulate-like (autholite) xenoliths are rare in the pyroclastic rocks and might indicate that the crystals accumulated in crustal magma chambers. The rock fragments vary in size between 4–5 cm to even N1 m. (iv) Trachytic and leucite–clinopyroxene tephritic dikes are injected in the previously emplaced volcanic (stage ii) and/or in pyroclastic units (stage iii). The dikes are variable in size, from 30–40 cm up to 2 m. Near the Saray village, a set of dikes with different composition are injected within the pyroclastic rocks. Rarely, the earlier dikes crosscut the older ones. (v) A late stage phase mainly accompanying the Lahar formation. (vi) Caldera collapse stage and formation of ring and radial lamprophyric dikes, cutting the laharon deposits near the Saray village. Lamprophyric sills and tephritic–basanitic dikes are found elsewhere within the laharon deposits, far from the Saray caldera. Moreover, clinopyroxene-rich tephritic lavas occur within the laharon deposits, in the absence of any sharp contacts, and with a red-colored paleosol.

Magmatic activity of stage ii is mainly characterized by at least five successive cycles. The first cycle is characterized by a pyroclastic sequence with leucite or analcimized leucite-bearing tephritic and/or leucititic (40–80% leucite) rock fragments. Trachytic lavas with large sanidine phenocrysts represent the second cycle followed by leucite–clinopyroxene–rich tephrites (third cycle). The fourth cycle is a combination of thick (70–80 m) leucite-bearing tephritic lavas and subsequent pyroclastic breccias. Dikes (even 30 m thick) are common within the leucite tephrites. Highly vesicular, leucite-bearing tephritic lavas rest over the pyroclastic rocks (fifth cycle). The contact of the rocks shows peperitic characteristic with red-colored bombs and pumiceous lavas.

3. Petrography of high-K volcanic rocks

About 60 thin sections from all rock units of the Saray potassium-rich rocks were examined in this study. From these, nine representative samples were selected for microprobe analyses, thirty samples for whole rock major and trace (REE) element analyses, two samples for Ar–Ar dating and 8 samples for Sr–Nd–Pb isotope studies (see Appendix 1 for analytical methods).

Nomenclature and classification of high-K rocks have been long debated (e.g., Foley et al., 1987; Mitchell and Bergman, 1991). The complex mineralogy of these rocks is attributed to the complex chemical compositions of their source magma(s) (e.g., Foley et al., 1987; Conticelli et al., 2012) and subsequent mixing between two types of magmas and/or assimilation of crustal materials. Here we adopted a classification based on field occurrence, petrographic features and mineral modes. Thus, we subdivided the rocks into seven groups including trachytic–phonolitic–tephritic lavas, trachytic dikes in pyroclastic sequences, rock fragments in volcaniclastic rocks, tephritic–basanitic dikes, leucitic lavas, lamprophyric dikes within lahar deposits and nepheline syenite xenoliths.

3.1. Trachytic–phonolitic–tephritic lavas (nepheline-bearing)

The trachytic lavas consist mainly of highly zoned, euhedral to rarely subhedral and green clinopyroxene phenocrysts (0.5–1 mm). Phlogopite (biotite) is surrounded by fine-grained Fe–oxide grains, which can be correlated to the breakdown of phlogopite during decompression and the oxidizing conditions during eruption. Fine-grained (0.2–0.3 mm) microphenocrysts of nepheline occur rarely. Apatite and zircon occur as tiny inclusions within pyroxene. The groundmass is composed of sanidine and plagioclase (with anorthoclase) associated with clinopyroxene and phlogopite microlites (Fig. 3B). Oxide minerals are common within the groundmass. Greenish sodic amphiboles occur around the clinopyroxene grains. Leucite and analcimized leucite (>0.5 mm but av. = 0.2–0.3 mm) associated with apatite and clinopyroxene with a potassium feldspathic groundmass is widespread in the phonolitic–tephritic lavas (Fig. 3A and F). Euhedral to subhedral, olivine phenocrysts (<0.5 mm but rarely 2 mm) are abundant in basanites. Ocelli texture is common in the fine-grained groundmass of some basanites. Two types of clinopyroxenes are common within the tephrites: 1 – green and sodic clinopyroxene, subhedral to euhedral with sieve texture (0.5–1 mm) and 2 – large (>1 mm) clinopyroxene with pale-green or colorless core and green rim. The interstices in the trachytic lavas are filled by perthitic alkali feldspar.

3.2. Trachytic dikes in pyroclastic units

These rocks are porphyritic, with clinopyroxene and phlogopite microphenocrysts. Sanidine microlites, oxide minerals and fine-grained clinopyroxene and phlogopite crystals are abundant in the groundmass. Trachytic dikes have two types of clinopyroxenes; pale-green and deep green varieties (Fig. 3E). Titanite is common in some of these dikes.

3.3. Rock fragments in volcaniclastic rocks

The rock fragments within volcanic breccias show different modal compositions and can be subdivided into quartz monzonite to quartz monzo-syenite fragments, nepheline-bearing monzonites, tephrites and tephritic tuffs. Euhedral to subhedral plagioclase phenocrysts (1–2 mm) with alteration into clay and sericite are abundant in the quartz-monzonite to quartz monzo-syenitic rock fragments. Anhedral and fine-grained quartz crystals (~20 vol.%) are present between plagioclase laths and accompany the K-feldspar grains. K-feldspar occurs as interstitial phase between other minerals. Deep-green clinopyroxene and apatite microphenocrysts make the other rock components. Nepheline (~2 vol.%) associated with sodic clinopyroxenes, plagioclase, K-feldspar, phlogopite and amphibole occur in nepheline bearing monzonites and/or monzo-syenites. The tephritic tuffs contain lithic fragments and altered glass shards. Olivine, green and/or colorless clinopyroxenes and K-feldspar crystal fragments are abundant.

3.4. Tephritic–basanitic dikes in volcanic and/or pyroclastic rocks

Zoned large (3–4 mm) and medium (1 mm) clinopyroxene grains are the major components in these rocks. The core of some large-sized clinopyroxenes exhibits a pale-green color while their rim shows a deep green color. The clinopyroxenes locally exhibit sieved-textured domains surrounding the clear core (Figs. 3D and 4C). The evolution of the Sieve texture is widely attributed to changes in temperature, pressure and composition in the magma (e.g., Tsuchiyama, 1985; Nelson and Montana, 1992). Euhedral to subhedral olivine (1 mm to <1 mm) is common, showing slight alteration into iddingsite (Fig. 3D). Large to small grains of leucite phenocrysts (or analcimized leucite) (<1 mm) are abundant (Fig. 3C). Alkali-feldspar overgrowth around the leucite is also noticed. Nepheline microphenocrysts are rare. The groundmass of these rocks is holocrystalline, composed of K-feldspar, clinopyroxene, oxide minerals and fine-grained, highly altered olivine.
crystals. There are two generations of clinopyroxene in most of the dikes; 1 — colorless clinopyroxenes, sometimes with a pale green, thin rim; 2 — green clinopyroxene. The occurrence of these two clinopyroxene types might indicate mixing between alkaline and crustal-derived magmas. Most of the pyroxenes exhibit sieve texture, again suggesting a magma mixing process, although this feature could also reflect rapid decompression and volatile loss in a magma that has not been mixed.

3.5. Leucite lavas

These rocks are represented by the occurrence of subhedral to euhedral, large leucite phenocrysts (av. = 0.5 mm up to >4 mm) and large (av. = 0.5–1 mm up to >5 mm) zoned sodic clinopyroxenes. Leucite constitutes >60 vol.% of these rocks. Olivine, apatite (0.2–0.3 mm) and oxide minerals (<0.1 mm) microphenocrysts are common. Some clinopyroxenes exhibit resorbed texture. Leucites have clinopyroxene inclusions. The groundmass of the rocks is holocrystalline with clinopyroxene and leucite fine-grained crystals.

3.6. Nepheline syenite xenoliths

Xenoliths contain very large (>6 mm but mostly 2–4 mm) nepheline phenocrysts. Deep-green, sodic clinopyroxenes, phlogopite and albite are present between the nepheline crystals (Fig. 4D). Biotite and clinopyroxene are elongated prismatic in shape with width/length ratio of around 1/7 to 1/20 for biotite and 1/5 to 1/10 for clinopyroxene. Brown amphiboles are present around the clinopyroxene. Sphene occurs as euhedral and primary (magmatic) crystals. Apatite, clinopyroxene and phlogopite are included within nepheline crystals.
3.7. Lamprophyric (sannaite) dikes within lahar deposits

Olivine, clinopyroxene and phlogopite are the main constituents of the lamprophyric dikes, set in a groundmass composed of altered alkali-feldspar and sodic clinopyroxene microlites. Clinopyroxenes are either pale-green and zoned (0.5–1 mm) or colorless (0.2–0.3 mm).

4. Mineral, whole rock and Sr–Nd–Pb isotopes geochemistry

4.1. Mineral chemistry

4.1.1. Clinopyroxene

Following the classification of Morimoto et al. (1998), all clinopyroxenes of the K-rich rocks including the tephritic rocks fragments in volcaniclastic rocks, phonolitic–trachytic lavas, basanitic dikes, trachytics as well as sannaite dikes are diopsides (Fig. 5A) with variable $X_{Mg}$ values (molar Mg/Mg + Fe$^{2+}$) of 0.61 – 0.93. The Al and Ti contents of clinopyroxenes are highly variable, ranging from 0.05 to 0.37 and 0.01 to 0.06 a.f.u. (Appendix 2a). In the clinopyroxenes from trachytic–phonolitic–tephritic (TPT) lavas, Al is not sufficient to fill completely the tetrahedral site, similar to clinopyroxenes from Mediterranean lamproites (e.g., Cellai et al., 1994; Conticelli et al., 2012) (Fig. 5B). Most of the clinopyroxenes from trachytic–basanitic dikes have higher Al and Ti contents similar to transitional high-K rocks of Conticelli et al. (1997)(Fig. 5B).

Clinopyroxene phenocrysts of the Saray volcanic rocks display oscillatory zoning, with high Ti and Al cores, mantled by low Ti and high Al clinopyroxenes, and grading into low Ti and Al outer rims, and/or vice versa, with poor Al–Ti core rimmed by high Ti–Al clinopyroxenes (Fig. 4A,B). The older clinopyroxenes typically show complex resorbed grain margins with a rim of high Ti–Al clinopyroxene (Fig. 4C).

The rim shows a sieve texture, denoting decompression during the rapid ascent of the magma to the surface. The Cr content of the clinopyroxenes is relatively low (Cr$_2$O$_3$ b 0.5 wt.%) (Appendix 2a).

4.1.2. Feldspar

Plagioclase is rare in the high-K rocks, except as a minor component in the tephritic rock fragments and shows labradorite composition ($An_{53}$) (Fig. 5C). Potassium feldspar microlites of the Saray volcanic rocks have a fairly large compositional range between $Or_{15}$ and $Or_{98}$. The alkali feldspars in tephritic rock fragments have anorthoclase composition ($An_{11–19}Ab_{60–97}Or_{15–29}$) (Appendix 2d). The Fe content of feldspar grains varies between 0.12 and 0.86 wt.% (Appendix 2d).

4.1.3. Mica

The mica in the Saray volcanic rocks has biotite to phlogopite composition with $X_{Mg}$ ranging from 0.58 to 0.89. Their Al$_2$O$_3$ is nearly constant with limited variation between 11.3 and 13.5 wt.%. These micas plot in the alkaline and calc-alkaline lamprophyric fields of Rock's (1987) diagram (Fig. 6A). The TiO$_2$ and Cr$_2$O$_3$ abundances of the Saray micas vary from 1.4 to 8.2 and 0 to 0.6 wt.% respectively. The mica composition seems to be different in various types of the Saray lavas including high Ti but low Fe and high Fe and low Ti varieties (Appendix 2b). Some of the micas are high Ti phlogopites, and the charge balance with Fe$^{3+}$ might require a tetraferriphlogopite component — i.e., Fe$^{3+}$ in the tetrahedral site.

4.1.4. Opaque minerals

Opaque minerals in the Saray volcanic rocks are generally represented by magnetite and Ti–magnetite (Fig. 6B). These grains are...
characterized by highly variable contents of MgO (0.04–7.8 wt.%) and TiO$_2$ (0.36–23.61 wt.%) (Appendix 2c).

4.1.5. Leucite and analcimized leucite

Leucite phenocrysts are characterized by high contents of K$_2$O (20–22 wt.%) and Al$_2$O$_3$ (22–24 wt.%) whereas the analcimized leucites have lower K$_2$O (9–14 wt.%) and Al$_2$O$_3$ (19–23 wt.%) (Appendix 2d, f), but with higher water content. Na$_2$O abundance of leucite and analcimized leucite varies from 0.02 to 0.4 and 0.1 to 3.8 wt.% respectively.

4.1.6. Apatite

Apatite occurs as microphenocrysts in the Saray high-K rocks with low levels of FeO (0.1–0.8 wt.%) (Appendix 2g).

4.1.7. Olivine

Olivine in the Saray high-K rocks varies in composition from Fo$_{76}$ to Fo$_{91}$, mostly from the core of polygonal to resorbed crystals. NiO shows positive correlation with forsterite content whereas CaO shows a negative correlation (Fig. 6C,D), with NiO reaching values as high as 0.2 wt.% (Appendix 2h), different from those of the lamproitic rocks. Olivine displays a decrease of CaO with increasing forsterite, in contrast.
to the trend suggested for lamproitic rocks by Conticelli and Peccerillo (1992), Conticelli et al. (2012) and Prelević and Foley (2007).

4.1.8. Nepheline

Nephelines in these rocks have high contents of Na$_2$O (15.4–16.3 wt.%), Al$_2$O$_3$ (29.7–31.2 wt.%) and low contents of K$_2$O (2.8–3 wt.%) (Appendix 2j).

4.2. Whole rock geochemistry

Major and trace (REE) element analyses of representative high-K samples are listed in Appendix 3. The LOI of the Saray high-K rocks vary between 0.2 and 6.1 wt.% but most of them (22 samples) have less than 2 wt.% LOI. High LOI values are common in the samples with analcimized leucites and altered glass groundmass. Owing to the relatively high LOI values of some samples, all major elements have been recalculated to 100% on the water-free basis for plots in variation diagrams.

The Saray volcanic rocks (and nepheline syenite xenoliths) were classified using the total alkalis (K$_2$O + Na$_2$O wt.%) vs. SiO$_2$ diagram of Le Bas et al. (1986) (Fig. 7A). The high-K rocks fall in a wide range represented by tephrite to phonolite/trachyte and trachy-andesite fields, in accordance with their petrographic features. All Saray volcanic rocks lie within shoshonite to ultrapotassic fields defined by Peccerillo and Taylor (1976) (Fig. 7B). These volcanic rocks display a low to high potassic affinity (K$_2$O/Na$_2$O) with K$_2$O/Na$_2$O ratios ranging from 0.6 to 4.5 (Appendix 3).

In the CaO vs. Al$_2$O$_3$ and MgO diagrams of Foley et al. (1987), most of the Saray high-K rocks display affinity to Roman-type ultrapotassic rocks (Fig. 7C,D). These rocks are different from normal lamproites.
NW Iran lavas modified after Le Bas et al. (1986).

The Turkish high-Mg mantle-derived ultrapotassic rocks of lamproitic affinity in the classification diagrams of Foley et al. (1987).

The increase in Al$_2$O$_3$ and TiO$_2$ might suggest accumulation of feldspar (sanidine) in the more acidic rocks (trachytic rocks). The scatter of Na$_2$O and K$_2$O shows only minor variation from 9 to 13 wt.%, where the highest values indicate clinopyroxene accumulation.

The behavior of trace elements vs. silica is markedly different from that of the major elements (Fig. 8). All the samples show relatively similar Zr abundance (179–396 ppm) and in Zr vs. SiO$_2$ diagram, the rocks display a horizontal trend. The trachytic dikes/lavas have higher Zr content (656–805 ppm) (Appendix 3). The same trend is also clear for Nb values and the Nb abundance is N74 ppm in trachytic lavas/dikes and b45 ppm in other rock types (Appendix 3). This duality in Nb–Zr distributions might suggest a similar origin for all rocks types displaying horizontal trend except for the trachytic rocks which possess higher Zr–Nb values, reflected in the modal abundance of amphibole, phlogopite, zircon and apatite. Although Rb and Th show scatter, they show an overall increase with SiO$_2$ (Fig. 8). The trachytic rocks also possess higher Th content, and might reflect the incompatibility of this element resulting in accumulation in the fractionated melt (glass). La and Y show a general negative trend with SiO$_2$ but with much scatter. In general, amphibole and phlogopite are absent in silica undersaturated (leucite-bearing) ultrapotassic lavas, but present in silica saturated trachytic rocks. The shift from silica-undersaturated to silica-saturated rocks could be interpreted by early separation of more silica-undersaturated minerals like olivine and clinopyroxene and by the early crystallization of leucite. This feature associated with scattered trend in Harker diagrams (and/or linear trends) for some elements can
be interpreted also by contamination by crustal materials and/or mixing with compositionally different melt (Sorensen et al., 1999).

In contrast to major elements, REEs are more dependent on the mineralogy of the mantle source and hence reflect the process of melt formation and extraction (Hofmann and Jochum, 1996). Although the lavas are characterized by various modal mineralogy and relatively different signatures of major and some trace elements, they display similar chondrite-normalized REE patterns (Fig. 9). The rocks are all strongly enriched in light REE (136–474 times chondrite for La) relative to HREE (8–21 times chondrite for Yb; Fig. 9). The La(0)/Yb(0) ratios vary between 13 and 40 for the Saray volcanic rocks and nepheline syenitic xenoliths. A high LREE/HREE ratio in the Saray high K-rocks could be related to the low degree of partial melting of an enriched mantle source. Partial melting of a garnet (−phlogopite) bearing mantle source could enrich LREE abundances relative to HREE (e.g., Zhang et al., 2008). Most of the Saray samples (although not all) have steep patterns from Ho to Yb, which might suggest the presence of residual garnet in the source of these magmas.

In primitive mantle-normalized diagram (Fig. 9), all the Saray high-K rocks are mainly characterized by strong enrichment in LILEs (e.g., Rb, Ba, U, K, and Th) relative to La. The lavas are characterized by conspicuous negative anomalies of HFSE including Nb, Ta and Ti relative to LREEs (Fig. 9). Phosphorous shows depletion in apatite-poor samples whereas a positive anomaly is displayed in apatite-bearing samples. Enrichment in LILEs and depletion in HFSEs are considered as main features of subduction-zone-related lavas (e.g., Pearce and Cann, 1973; Pearce, 1982; Pearce and Peate, 1995).

4.3. Sr–Nd–Pb isotope geochemistry

The Nd–Sr–Pb isotopic ratios of the Saray volcanic rocks are given in Table 1. The initial (11 Ma) 87Sr/86Sr ratio of these rocks varies between 0.70662 and 0.70797. Their initial 143Nd/144Nd ratio shows a narrow range between 0.51242 and 0.51252. All the Saray high-K volcanic rocks plot within the enriched quadrant of the conventional Nd–Sr isotope diagram (Fig. 10). The lavas show a tendency to EM2 OIB-mantle source with affinity to lamprophyres of the Alps (Fig. 10). The rocks show a distinct continental signature rather than the features of MORB and the Late Cretaceous ophiolitic lavas. The EM2 mantle signature is suggested to be inherited from recycling slab sediments (Zindler and Hart, 1986), although the origin of the EM2 signature could also involve metasomatism (i.e. melt/fluid infiltration) of both oceanic and continental lithosphere followed by recycling and long-term storage of this lithosphere (Workman et al., 2004).

The 206Pb/204Pb and 208Pb/204Pb values vary between 19.09 and 19.26 and 39.27 and 39.46 respectively (Table 1). The samples are characterized by highly radiogenic 207Pb/204Pb ratios, ranging from 15.69 to 15.70 (Fig. 10B). The lavas plot above the Northern Hemisphere Reference Line (NHRL; Hart, 1984), in both 207Pb/204Pb and 208Pb/204Pb diagrams (Fig. 10B,C). The highest 206Pb/204Pb and 208Pb/204Pb values are markedly higher than GLOSS and EM2 reservoir values (although 208Pb/204Pb ratio is between GLOSS and EM2) and indicate the involvement of subducted terrigenous sediments and/or continental crust in the source. These values could also correspond to an older reservoir of high-integrated Th/U.

5. Ar–Ar geochronology

The two sets of Ar–Ar age data yielded comparable ages on biotite grains for the Saray high-K trachytic lavas. The results are presented in Table 2 and plotted in Fig. 11. The 40Ar–39Ar data are presented in Appendix 5 together with a detailed discussion of each age. Laser step-heating experiments on biotite concentrates for two volcanic samples (E11-3 and E11-18) show the same ages.
E11-3 displays a well-defined plateau age of 10.3 ± 0.10 Ma. This age is indistinguishable from the isochron age of 10.3 ± 0.17 Ma (Table 2). The best age estimate of 10.9 ± 0.13 Ma for sample E11-18 comes from the plateau age which is comparable with the Ar–Ar isochron age of 10.7 ± 0.25 Ma.

6. Discussion

The origin and evolution of magmatic rocks along convergent plate margins involve a series of processes (e.g., Eyuboglu, 2013) including metasomatism of the mantle wedge by melt and/or fluids released by the downgoing slab (e.g., Pearce, 1982; Hawkesworth et al., 1991; Kelemen et al., 2004; Khedr et al., 2010), variable degree of partial melting of the metasomatized mantle (e.g., Pearce and Parkinson, 1993) and subsequent process at shallower depths such as magma mixing, AFC (e.g., DePaolo, 1981) and/or combinations of these processes (Ersoy et al., 2010). In this section we will briefly discuss the role of each process in the genesis of the high-K Saray lavas.

6.1. Subduction fingerprints of the Saray high-K volcanic rocks; arc-related setting?

The Saray high-K volcanic rocks are characterized by troughs in Nb, Ta, Ti and Zr (except the trachytes) and strong enrichment in incompatible elements including Rb, Ba, Th, U, K and LREE in a primary mantle-normalized diagram (Fig. 9). The enrichment in incompatible elements implies that the melt source was a metasomatized sub-continental lithospheric mantle (SCLM). The negative anomalies in Nb, Ta and Ti are typical features of subduction-related magmatism. A relatively high Sr/Y ratio, but also high Y content (resulting from clinopyroxene accumulation in most samples) is characteristic of the Saray volcanic rocks, different both from adakites and arc-like rocks (Fig. 12A). The influence of subduction components on the mantle source of the Saray volcanic rocks can also be detected in a Th/Yb vs. Ta/Yb diagram (Pearce, 1982) (Fig. 12B), where the rocks show high Th/Yb content (tending towards the value of average continental crust). This feature might suggest both subduction influence and/or crustal contamination of the lavas (AFC trend). The higher Th/Yb ratios also could indicate a lithospheric mantle source enriched by subduction components (Dilek et al., 2010). On the other hand, trachytic lavas have higher Th/Yb and Ta/Yb ratios than those of the other primary magmas, indicating that fractional crystallization also played an important role during the generation of the rocks. The effect of fractional crystallization can be traced in Th/Yb vs. SiO diagram (Fig. 12C), where the fractionated lavas (with higher SiO2) have a higher Th/Yb ratio. The high LILE and LREE contents of the Saray lavas associated with radiogenic 207Pb/204Pb ratio are also similar to those of lavas formed in post-collisional setting (e.g., Turner et al., 1996; Zhao et al., 2009; Dilek et al., 2010).

There is a subtle but clear positive Nd anomaly in virtually all of these rocks, although this is not reflected in the REE patterns of most the alkalic rocks. The magmas that consistently show positive Nd anomalies are kimberlites and low-SiO2 lamproites, including most leucite-bearing rocks. This feature could suggest that the parent magma(s) were not derived from a single-stage melting or crystallization process, but most likely reflects an early enrichment event when the LREEs were strongly enriched, followed by a melting event resulting in the selective removal of La, Ce and Pr. The distribution coefficients for the REE that define the Nd anomalies require garnet in the magma source. The depletion of Nb, Ta, and Ti clearly suggests that the dominant source of these magmas has been modified by subduction-related processes, or that a phase like rutile, that preferentially retains Nb, Ta, and Ti, is stable in the melt source. In the studied rocks, some trachytic rocks have negative P anomalies, whereas this anomaly is absent in the other rocks. This feature is apparently related to the apatite distribution, although the reasons remain vague.
6.2. Petrogenetic model for the formation of high-K rocks: combination of various hypotheses

Chondrite- and primitive mantle-normalized REE and multi-element (Fig. 10), as well as Sr–Nd–Pb isotope (Fig. 10) diagrams indicate that the source region(s) of all these volcanic rocks is similar. The lavas have also the same ages as revealed by Ar–Ar dating results (Fig. 11). However, complex zoning of the clinopyroxene and other minerals in the Saray lavas associated with their major, trace (REE) and Sr–Nd–Pb isotopes suggest that different processes including fractional crystallization,
assimilation–fractional crystallization and/or magma mixing could have played important roles during their formation.

### 6.2.1. Fractional crystallization: subordinate mechanism for the evolution of the Saray rocks?

The trends displayed by major and trace elements in Harker diagrams (Fig. 8) suggest that fractional crystallization was predominant during the formation of the Saray high-K rocks. Whereas Zr should show enrichment during fractionation (with increasing SiO$_2$) in the present case, all the samples display relatively similar Zr abundance except for trachytic dikes/lavas that have high Zr content. This is also true for Nb. Y shows lower values with highest SiO$_2$ for trachytic rocks which might reflect clinopyroxene (and amphibole) fractionation. The other samples record clinopyroxene accumulation with higher Y content. All the samples display negative Eu anomalies. Because most of the samples in this study do not contain amphibole (the only other phase that combined with apatite, could potentially produce Eu anomalies), the negative Eu anomaly probably derives from fractional crystallization of feldspars, and this indicates that none of the rocks represents a “primary” magma composition that reflects a melt from a mantle source. The nature of the negative Eu anomalies suggests that the rocks evolved by fractional crystallization, and did not originate from partial melting in the presence of feldspar in the source. For most mafic alkaline rocks, the pressure of melting is probably above the stability range of feldspars (~10 kb), so that the Eu anomaly is unlikely to be formed during partial melting. As mentioned earlier, the shift from silica-undersaturated to silica-saturated rocks could be also interpreted by early separation of more silica-undersaturated minerals such as olivine and clinopyroxene and by early crystallization of leucite. The effect of fractional crystallization is also displayed in a Th/Yb vs. SiO$_2$ diagram (Fig. 12C).

The major elements define an apparent liquid line of descent, although most of the incompatible trace elements do not show the behavior of a typical fractionating system. Thus, fractional crystallization alone cannot explain the other features of these rocks including the complex oscillatory zoning of pyroxene and/or their isotopic characteristics. Other mechanisms such as AFC and/or magma mixing should also be considered.

### 6.2.2. Assimilation-fractional crystallization processes (AFC): an alternate mechanism?

Assimilation and fractional crystallization processes are common in most Anatolian high-K rocks (e.g., Gulec, 1991; Aldanmaz et al., 2000; Yilmaz et al., 2001; Dilek et al., 2010; Ersoy et al., 2010; Eyuboglu, 2010). Th/Yb vs. Ta/Yb relationships of the Saray lavas and their radiogenic $^{206}$Pb/$^{204}$Pb ratio also show the effect of both fractional crystallization (FC) and assimilation-fractional crystallization (AFC) processes. In order to evaluate the effect of AFC, we used the FC–AFC–FCA and mixing modeler software of Ersoy and Helvacı (2010). Sample E11-29 having the lowest SiO$_2$ and relatively high MgO (with both low LOI and $^{87}$Sr/$^{86}$Sr ratio) is assumed to represent the starting magma composition for AFC modeling.

An average fractionating composition of plagioclase (65%) + clinopyroxene (15%) + amphibole (10%) + biotite (10%) following Ersoy et al. (2012a) is used (Fig. 13), except in Fig. 13E where the starting composition used was that of sample E11-58 with fractionating mineral assemblage of plagioclase (40%) + clinopyroxene (30%) + amphibole (10%) + biotite (10%) + olivine (10%). These considerations also fit well with our observations in the field showing clinopyroxene–plagiolepsis–feldspar (colivine) cumulate-like assemblages in pyroclastic deposits. The metamorphic rocks (granitic gneiss; sample CHJ09-12) of the Chah-Jam complex from central Iran with Cadomian age (ca. 550 Ma; Shafaii Moghadam et al., unpublished data), are considered as assimilant in the mid-crustal level, exhumed during the extension phases in the late Cretaceous in Iran (Hassanzadeh et al., 2008). These rocks have the same ages and similar mineral assemblages to those of the Cadomian granitic gneisses that host the high-K Saray volcanic rocks. The partition coefficients and the composition of starting
material and assimilant in FC and AFC models are given in Tables 3 and 4. The interaction of the starting melt with the assimilant will result in an increase of $^{87}\text{Sr}/^{86}\text{Sr}$ and decrease in $^{143}\text{Nd}/^{144}\text{Nd}$ of the contaminated magma (Pb values are not available in the case of assimilant; sample CHJ09-12, to model the AFC using Pb isotopes). To test the effect of contamination of the Saray lavas with mid-crustal rocks (as evidenced by Chah-Jam metamorphic rocks; CJMR), the $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the volcanic rocks are plotted against Th and Sr and also vs. the Sr/Rb ratio (Fig. 13). The trend of the AFC model best explains the $^{143}\text{Nd}/^{144}\text{Nd}$ vs. Sr/Rb and Sr compositions of the Saray volcanic rocks using the CJMR as assimilant and low ($r = 0.2$) to medium ($r = 0.4$) degrees of assimilation, and the trends displayed by $^{87}\text{Sr}/^{86}\text{Sr}$ ratio vs. Th and Sr. As minor olivine is present in the Saray primitive lavas (and in autholiths within the pyroclastic rocks), we consider another fractionating mineral assemblage of plagioclase (40%) + clinopyroxene (30%) + amphibole (10%) + biotite (10%) + olivine (10%) (with CJMR as assimilant) and using V/Y vs. Th to test the contamination effect. The strongest positive correlation is achieved by this model and using ratios of assimilation to crystallization ($r$) from 0.2 to 0.8 (Fig. 13E). Although the effects of contamination can be traced among the Saray volcanic rocks, especially in the case of the V/Y vs. Th plot, the petrographic and geochemical features of the lavas such as the presence of both Ti–Al rich and Ti–Al poor Cpx phenocrysts in same rocks, complex zoning of the clinopyroxenes (and other phases), high $^{87}\text{Sr}/^{86}\text{Sr}$ values, and the high abundances of LREEs and LILEs cannot be explained either by FC or by AFC processes. In addition, the content of incompatible trace elements in the Saray rocks is higher than that of crustal end member, suggesting that crustal contamination would have diluted rather than enriched the trace element content of the Saray magmas. Therefore other factors including mantle source heterogeneity due to the metasomatic events relating to sediment fluids/melts from descending slab and/or different degrees of partial melting of the metasomatized mantle should also be considered.

6.2.3. Source heterogeneities and mixing; a more favorite mechanism?

Post-collisional magmatic rocks with shoshonitic and ultra-potassic (with high K and/or high MgO and with different degree of silica saturation/undersaturation) signatures are abundant in the Mediterranean–Iran regions, and their origin has been correlated to partial melting of clinopyroxene–amphibole–plagiopitite veins in a metasomatized mantle source (e.g., Gulec, 1991; Prelević et al., 2004, 2005, 2008; Ersoy et al., 2008; Conticelli et al., 2009; Pe-Piper et al., 2009; Conticelli et al., 2012 and references therein). Their genesis is suggested to be related to recycling of pelitic sediments and subsequent K-metasomatism and enrichment of peridotic lithosphere (e.g., Conticelli et al., 2012). Factors that control the silica activity of the high-K melts are suggested to be related to: 1 — abrupt change of the comp-position of the recycled sediments (e.g., Conticelli and Peccerillo, 1992; Frezzotti et al., 2009) and d/o r 2 — interaction between deep asthenospheric mantle with lithospheric portion of mantle wedge through slab tears within the descending slab (e.g., Conticelli et al., 2012; Prelević et al., 2012). From petrographic and geochemical evidence, it is clear that some mixing between newly supplied silica under-saturated magma and batches of pre-existing calc-alkaline magmas might have occurred during the evolution of the Saray volcano.

At convergent margins, the composition of the mantle wedge can be modified by the addition of subduction-related components such as melts and/or fluids released from the subducting slab (sediments and/or underlining MORB-type basalts) (e.g., McCulloch and Gamble, 1991; Hawkesworth et al., 1997). Some elements are highly mobile in fluids such as LILEs whereas others such as Th and LREEs are enriched in sediments and can be concentrated in sediment-derived melts. We attempted to model the effects of sediment fluids (SF) and sediment melts (SM) on the mantle wedge (primary mantle; PM) using ratios like as Th/Nb vs. Ba/Nb and Ce/Yb vs. Sr/Yb (Fig. 14A,B). The Saray high-K rocks are characterized by elevated Th/Nb but variable Ba/Nb ratios. A higher Ba/Nb ratio could be interpreted as evidence for higher degree of sediment-derived fluid contribution rather than sediment melts (if we neglect the probable Ba mobility during alteration). However the Saray rocks have elevated Ce/Yb and Sr/Yb ratios, mostly compatible with sediment melt (SM) components. Higher Sr/Yb and Ce/Yb could also reflect the effect of residual garnet in the mantle source (Esoy et al., 2010). The presence of a sediment-derived, subduction-related metasomatic component with high SM/La and Th/La ratios (i.e., SALATHO; Tommasini et al., 2011) is suggested to be responsible for formation of the Mediterranean lamproites (e.g., Conticelli et al., 2011; Tommasini et al., 2011; Conticelli et al., 2012). However in the Saray volcano, the trachytic dikes have a higher Th/La ratio and indeed, these rocks are compositionally different from lamproites by having lower Mg# and higher Al$_2$O$_3$. The SALATHO component is interpreted to derive from melting of accreted mélangé during the collision of the Gondwana-derived continental blocks with Eurasia associated with the closure of Neotethyan oceanic basins (Tommasini et al., 2011).

Small mantle wedge heterogeneity, as described for Mediterranean high-K rocks (e.g., Conticelli and Peccerillo, 1990; Foley, 1992; Conticelli et al., 2012), involving the interaction between calc-alkaline and lamproite-like (ultrapotassic) components could be responsible for the genesis of the Saray lavas. In this model, the ultrapotassic component is related to a metasomatized mantle with vein networks whereas the calc-alkaline component is derived from the mantle surrounding the vein networks (e.g., Foley, 1992; Perini et al., 2004; Bianchini et al., 2010). Both components are present in the same lithospheric mantle. The calc-alkaline and ultrapotassic melts responsible for the Saray high-K rocks could have been produced at the same time from different mantle source levels and mixed in upper levels within crustal magma chambers. The presence of metasomatized mantle with vein networks reinforces the chemical effects of SM and SF addition on the PM-like mantle source. In order to estimate the amount of the subducted sediment contribution to the mantle source of the Saray volcanic rocks, we used their Sr–Nd isotopic ratio, compared with other end-members including primary mantle and subduction components (Fig. 14C,D). In this diagram, the mixed mantle source (MMS) is obtained from 10% mixing of PM with a subducted sediment component (Esoy et al., 2010). On the Sr–Nd isotope diagram (Fig. 14C), the Saray volcanic rocks define a binary mixing trend, representing 5–8% mixing between the primary mantle (PM) and sediment melts. A sedimentary component in late Miocene Saray volcanic rocks is also evidenced from Pb isotopes data of the lavas (Fig. 10). Both Fig. 10A and B suggest that this is the contribution from an old source. Particularly, the elevated $^{207}$Pb/$^{204}$Pb requires an old source (i.e., a source from where the decay of $^{235}$U generated high $^{207}$Pb, and which was subsequently isolated from mixing with the younger component). The Pb isotope data in Fig. 10B are consistent with similar data from Turkey (e.g., Prelević et al., 2012), showing compositions that are offset to the high side of NHRL and this is a regional pattern. The highly radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{207}$Pb/$^{204}$Pb values for the Saray lavas are also another robust evidence for the involvement of slab sediments. Melting of subducted sediments might have supplied the crustal components to the mantle source of the Saray volcanic rocks (e.g., Prelević et al., 2005, 2012; Conticelli et al., 2012). The Saray lavas, like most of the Mediterranean lamproites, plot in the mixing trend between melts derived both from high Nb/U as well as less radiogenic Sr source (mantle component 2) and melts generated from a mantle contaminated by subducted sediments and/or crust-derived sedimentary materials with low Nb/U and radiogenic Sr isotope (mantle component 1; Fig. 14D) (Prelević et al., 2012). Prelević et al. (2012) suggested that this addition should be as sediment partial melt patches, evidenced by high Th/Nb and/or Th/Yb ratios. The mantle component 1 is characterized by highly radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{207}$Pb/$^{204}$Pb, non-radiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ and a relatively non-radiogenic $^{206}$Pb/$^{204}$Pb. The continental crustal component in the Saray high-K rocks is inferred to have originated from turbiditic sediments as indicated by the very high $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{207}$Pb/$^{204}$Pb. The $^{207}$Pb/$^{204}$Pb, $^{206}$Pb/$^{204}$Pb
and $^{206}\text{Pb}/^{204}\text{Pb}$ isotopes ratios of the Saray lavas are higher than those of global subducting sediments (GLOSS) and enriched mantle II (EM2). The Saray samples have a (near to-) super chondritic Hf/Sm ratio, indicating that the subducted sediments had terrestrial origin. This is also well constrained from the $^{207}\text{Pb}/^{204}\text{Pb}$ ratio. The lavas might thus reflect the involvement of subducted terrestrial sediments and/or continental crust in the mantle sources, which have either high Nd/Hf (pelagic sediment) or low Nd/Hf (terrestrial sediments and/or continental crust) (e.g., Gasparon and Vare, 1998; Plank and Langmuir, 1998; Carpentier et al., 2009; Handely et al., 2011; Shafaii Moghadam et al., 2012).

### 6.2.4. Non-modal dynamic melting of metasomatized mantle source

Decompression melting of metasomatized lithospheric harzburgite with minor phlogopite and pargasite has been invoked for the formation of primary high-K magmas by dehydration melting at ~1 GPa, 1050–1150°C (Conceição and Green, 2004). In order to model the partial melting and $^{40}\text{Ar}/^{36}\text{Ar}$ ages for trachytic lavas.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock</th>
<th>Material dated</th>
<th>Plateau age (Ma)</th>
<th>Isochron age (Ma)</th>
<th>($^{40}\text{Ar}/^{36}\text{Ar}$)</th>
<th>MSWD $^{\text{a}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>E11-3</td>
<td>Trachytic lava</td>
<td>Biotite</td>
<td>10.33 ± 0.10 Ma</td>
<td>10.32 ± 0.17 Ma</td>
<td>312 ± 27</td>
<td>1.39/13</td>
</tr>
<tr>
<td>E11-18</td>
<td>Trachytic lava</td>
<td>Biotite</td>
<td>10.93 ± 0.13 Ma</td>
<td>10.68 ± 0.25 Ma</td>
<td>305 ± 8</td>
<td>0.42/9</td>
</tr>
</tbody>
</table>

$^{a}$ Initial ($^{40}\text{Ar}/^{36}\text{Ar}$) calculated from the intercept of the best straight line defined with the data in the ($^{36}\text{Ar}/^{40}\text{Ar}$) vs. ($^{39}\text{Ar}/^{40}\text{Ar}$) correlation diagram.

$^{b}$ Goodness of fit of the best straight line and number of points used.

Fig. 10. (A) $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios 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 display cluster close to the EM2. The Eastern and Western Alps and Tuscan lamprophyres and Roman Province fields are from Bell et al. (2013); Data for Neyriz lavas and Salzzevar ophiolite gabbros are from Shafaii Moghadam et al. (unpublished data). Note that Neyriz ophiolitic lavas have a shift towards higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio due to the reaction with seawater; (B) $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ and (C) $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams for the Saray lavas. GLOSS = globally subducting sediments (Plank and Langmuir, 1998); PHEM = primitive helium mantle (Farley et al., 1992).
trajectories, the mineralogical composition of the mantle source (associated with the partial melting degree of this mantle), from which the Saray volcanic rocks originated are inferred as shown in Fig. 15. Nond modal dynamic melting of a garnet-bearing amphibole lherzolite with Ulten Zone Peridotite (UZP), Eastern Alps (Rampone and Morten, 2001) composition (see Table 5) is modeled on Th/Yb vs. La/Yb and Tb/Yb diagrams (Fig. 15). The UZP is suggested to represent a paleo-mantle wedge formed above a subducted continental slab (Rampone and Morten, 2001). On this plot, the partial melting trajectories of UZP with variable garnet (5%, 7% and 10%) are shown. In Th/Yb vs. La/Yb diagram, the samples do not follow the melting trends of UZP mantle source whereas most of the primary lavas plot between 6 and ~8% partial melting of garnet lherzolite source of Walter (1998). In the Th/Yb vs. Th/Yb diagram, the primary samples do not follow the melting trajectory of the garnet herzolite (Walter, 1998). The samples have higher a Th/Yb ratio, probably resulting from the AFC process during ascent of the Saray melts (see Fig. 13E) and/or addition of high sediment melt input into the mantle wedge (Fig. 14A). The scatter of the samples in these diagrams seems to be the result of mixing between melts produced by different sources or by the same batch of melt passing through a compositionally-zoned mantle during multiple stages of partial melting and melt migration (Ersoy et al., 2010; Ersoy et al., 2012a,b). This interpretation is well constrained by the complex zoning of clinopyroxenes of the lavas. Trace (and REE) elements of the clinopyroxene from the Saray lavas in future studies could help to better understand this process.

Melting of phlogopite and apatite-bearing peridotites in the subcontinental lithospheric mantle is also proposed for the genesis of the Saray high-K rocks (Pang et al., 2013). However, our modeling (not shown) using both dolomite-bearing garnet peridotites (sample ULT18; Sapienza et al., 2009) and phlogopite-rich peridotites (MADRID; Gregoire et al., 2002 and/or Finero phlogopite harzburgite; Hartmann and Wedepohl, 1993) indicates that neither of these could produce the Saray ultrapotassic rocks. As phlogopite (and pargasite)-bearing lherzolites are suggested to be the source for the ultrapotassic lavas in the Anatolian regions (e.g., Prelević et al., 2012), it is possible that the phlogopite signature decreased through mixing of melts derived from different mantle sources (mantle components 1 and 2). This hypothesis is well constrained from complex oscillatory zoning of the Saray clinopyroxenes.

6.3. Geodynamic implications

Potassium-rich magmatic rocks occur widely in different tectonomagmatic environments such as continental active margins, postcollisional magmatic arcs and within plate settings (e.g., Muller et al., 1992; Mitchell et al., 1994; Elliott et al., 2010; Eyuboglu, 2010; Lu et al., 2012). It is widely accepted that the late Cenozoic magmatism in the Mediterranean–Iran regions occurred in a post-collisional setting and resulted from partial melting of a subduction-related metasomatized sub-continental lithospheric mantle (e.g., Aldanmaz et al., 2000; Alici Sen et al., 2004; Altunkaynak and Genç, 2008; Ersoy et al., 2010; Altunkaynak et al., 2012; Ersoy et al., 2012a,b; Prelević et al., 2012). Most of the high-K rocks from the Anatolian plateau are thought to have been derived from partial melting of a mantle source similar to UZP, either contaminated by multi-stages of oceanic subduction events during a long period and/or contaminated by continental subduction (Ersoy et al., 2010, 2012a,b). In the latter case, the deep subduction of upper continental crust might have led to metasomatism of the mantle by K-rich fluids (Schreyer et al., 1987). This process is mostly responsible for the genesis of lamproites. The Saray lavas are enriched in LILEs and LREEs relative to primary mantle and mark the presence of crustal components. This is also well constrained by more radiogenic 87Sr/86Sr and 206Pb/204Pb ratios for the Saray lavas. These features suggest that the Saray postcollisional late Miocene magmas were derived from small degrees of partial melting of subduction-metasomatized (subcontinental) lithospheric mantle source. However, mixing of melts from the two types of mantle source is suggested based on the isotope and trace elements ratios (e.g., Fig. 15). This scenario is well consistent with those of the other post-collisional magmatic rocks from the Miocene to Quaternary in the Turkish–Iranian High Plateau (e.g., Pearce et al., 1990; Keskin et al., 1998; Yilmaz et al., 1998; Dilek et al., 2010; Eyuboglu et al., 2012). Slab break-off and/or partial delamination (due to the roll-back of the subducting Tethyan lithosphere) of the
mantle lithosphere associated with partial melting of clinopyroxene–amphibole–garnet (and/or phlogopite) rich metasomatic veins in the lithospheric mantle in a post-collisional setting has been suggested as likely processes (Dilek et al., 2010). In this model, the first batch of lavas were derived mainly from the metasomatic veins and are mostly ultrapotassic as compared to the later high-K calc-alkaline and/or shoshonitic products which were suggested to be have resulted from melting of veins surrounding peridotites (Ersoy et al., 2010; Conticelli et al., 2012). However this model is not favorable for the Saray rocks as they have identical ages (~11 Ma) and nearly the same geochemical patterns. On the other hand, the mantle source of the Saray lavas is different from amphibole- (and/or phlogopite-) bearing tholeiites as...
### Table 3
Partition coefficients used in FC, AFC and melting models. (after Ersoy et al., 2010; Ersoy and Helvaci, 2006)

<table>
<thead>
<tr>
<th>Elements</th>
<th>Olivine</th>
<th>Opx</th>
<th>Cpx</th>
<th>Garnet</th>
<th>Amphibole</th>
<th>Biotite</th>
<th>Plagioclase</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cs</td>
<td>0.0004</td>
<td>0.01</td>
<td>0.13</td>
<td>0.003</td>
<td>0.046</td>
<td>0.626</td>
<td>0.13</td>
</tr>
<tr>
<td>Rb</td>
<td>0.002</td>
<td>0.003</td>
<td>0.0047</td>
<td>0.042</td>
<td>0.1</td>
<td>5.18</td>
<td>0.1</td>
</tr>
<tr>
<td>Ba</td>
<td>0.002</td>
<td>0.002</td>
<td>0.00061</td>
<td>0.023</td>
<td>0.45</td>
<td>3.48</td>
<td>0.3</td>
</tr>
<tr>
<td>Sr</td>
<td>0.00217</td>
<td>0.007</td>
<td>0.0963</td>
<td>0.012</td>
<td>0.45</td>
<td>0.183</td>
<td>2</td>
</tr>
<tr>
<td>Th</td>
<td>0.04</td>
<td>0.13</td>
<td>0.03</td>
<td>0.05</td>
<td>0.0145</td>
<td>0.01</td>
<td></td>
</tr>
<tr>
<td>U</td>
<td>0.045</td>
<td>0.035</td>
<td>0.04</td>
<td>0.015</td>
<td>0.08</td>
<td>0.0011</td>
<td>0.01</td>
</tr>
<tr>
<td>Zr</td>
<td>0.0045</td>
<td>0.03</td>
<td>0.121</td>
<td>0.3</td>
<td>0.4</td>
<td>0.0232</td>
<td>0.048</td>
</tr>
<tr>
<td>Hf</td>
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<td>0.055</td>
<td>0.263</td>
<td>0.3</td>
<td>0.5</td>
<td>0.19</td>
<td>0.051</td>
</tr>
<tr>
<td>Ti</td>
<td>0.011</td>
<td>0.024</td>
<td>0.1</td>
<td>0.1</td>
<td>0.69</td>
<td>0.9</td>
<td>0.04</td>
</tr>
<tr>
<td>Ta</td>
<td>0.000018</td>
<td>0.15</td>
<td>0.013</td>
<td>0.06</td>
<td>0.25</td>
<td>0.1069</td>
<td>0.04</td>
</tr>
<tr>
<td>Y</td>
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<td>0.438</td>
<td>2</td>
<td>1.1</td>
<td>0.007</td>
<td>0.03</td>
</tr>
<tr>
<td>Nb</td>
<td>0.0017</td>
<td>0.15</td>
<td>0.0027</td>
<td>0.1</td>
<td>0.3</td>
<td>0.0853</td>
<td>0.01</td>
</tr>
<tr>
<td>Sc</td>
<td>0.68</td>
<td>1.2</td>
<td>3.2</td>
<td>2.6</td>
<td>4.2</td>
<td>8.3</td>
<td>0.04</td>
</tr>
<tr>
<td>Cr</td>
<td>0.7</td>
<td>10</td>
<td>34</td>
<td>0.6</td>
<td>34</td>
<td>5.4</td>
<td>0.08</td>
</tr>
<tr>
<td>Ni</td>
<td>29</td>
<td>5</td>
<td>14</td>
<td>5.1</td>
<td>6.8</td>
<td>1.3</td>
<td>0.04</td>
</tr>
<tr>
<td>V</td>
<td>0.8</td>
<td>0.6</td>
<td>1.35</td>
<td>1.35</td>
<td>0.022</td>
<td></td>
<td></td>
</tr>
<tr>
<td>La</td>
<td>0.00064</td>
<td>0.002</td>
<td>0.0435</td>
<td>0.01</td>
<td>0.2</td>
<td>0.035</td>
<td>0.27</td>
</tr>
<tr>
<td>Ce</td>
<td>0.005</td>
<td>0.003</td>
<td>0.0843</td>
<td>0.021</td>
<td>0.35</td>
<td>0.03</td>
<td>0.2</td>
</tr>
<tr>
<td>Pr</td>
<td>0.0008</td>
<td>0.0048</td>
<td>0.124</td>
<td>0.054</td>
<td>0.35</td>
<td>0.035</td>
<td>0.17</td>
</tr>
<tr>
<td>Nd</td>
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<td>0.173</td>
<td>0.087</td>
<td>0.65</td>
<td>0.04</td>
<td>0.14</td>
</tr>
<tr>
<td>Sm</td>
<td>0.0013</td>
<td>0.01</td>
<td>0.283</td>
<td>0.217</td>
<td>0.85</td>
<td>0.03</td>
<td>0.11</td>
</tr>
<tr>
<td>Eu</td>
<td>0.0016</td>
<td>0.013</td>
<td>0.312</td>
<td>0.32</td>
<td>0.8</td>
<td>0.04</td>
<td>0.73</td>
</tr>
<tr>
<td>Gd</td>
<td>0.0015</td>
<td>0.016</td>
<td>0.336</td>
<td>0.498</td>
<td>0.95</td>
<td>0.03</td>
<td>0.066</td>
</tr>
<tr>
<td>Tb</td>
<td>0.0015</td>
<td>0.019</td>
<td>0.364</td>
<td>0.75</td>
<td>1.2</td>
<td>0.03</td>
<td>0.06</td>
</tr>
<tr>
<td>Dy</td>
<td>0.0017</td>
<td>0.022</td>
<td>0.363</td>
<td>1.06</td>
<td>1.3</td>
<td>0.035</td>
<td>0.055</td>
</tr>
<tr>
<td>Ho</td>
<td>0.0016</td>
<td>0.026</td>
<td>0.378</td>
<td>1.53</td>
<td>1</td>
<td>0.03</td>
<td>0.046</td>
</tr>
<tr>
<td>Er</td>
<td>0.0015</td>
<td>0.03</td>
<td>0.351</td>
<td>2</td>
<td>0.9</td>
<td>0.03</td>
<td>0.041</td>
</tr>
<tr>
<td>Tm</td>
<td>0.0015</td>
<td>0.04</td>
<td>0.297</td>
<td>3</td>
<td>0.9</td>
<td>0.03</td>
<td>0.036</td>
</tr>
<tr>
<td>Yb</td>
<td>0.0015</td>
<td>0.049</td>
<td>0.313</td>
<td>4.3</td>
<td>0.8</td>
<td>0.04</td>
<td>0.031</td>
</tr>
<tr>
<td>Lu</td>
<td>0.0015</td>
<td>0.06</td>
<td>0.265</td>
<td>5.5</td>
<td>0.8</td>
<td>0.03</td>
<td>0.025</td>
</tr>
</tbody>
</table>

### Table 4
The parameters used in FC, AFC, mixing and melting models (After Ersoy et al., 2010)

<table>
<thead>
<tr>
<th>Element</th>
<th>Primitive mantle (PM)</th>
<th>Sediment melt (SM)</th>
<th>Sediment fluid (SF)</th>
<th>Mixed mantle source (MMS)</th>
<th>Assimilant (CHJ09-12)</th>
<th>Starting melt (E11-29)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rb</td>
<td>0.605</td>
<td>89.46</td>
<td>28.67</td>
<td>6.45</td>
<td>104.50</td>
<td>149.50</td>
</tr>
<tr>
<td>Ba</td>
<td>0.750</td>
<td>1173.63</td>
<td>922.93</td>
<td>110.90</td>
<td>698.00</td>
<td>4410.00</td>
</tr>
<tr>
<td>Sr</td>
<td>20.300</td>
<td>436.56</td>
<td>411.37</td>
<td>63.67</td>
<td>140.50</td>
<td>1871.50</td>
</tr>
<tr>
<td>Th</td>
<td>0.083</td>
<td>8.69</td>
<td>1.44</td>
<td>0.58</td>
<td>11.55</td>
<td>32.10</td>
</tr>
<tr>
<td>Y</td>
<td>4.370</td>
<td>8.27</td>
<td>25.71</td>
<td>5.63</td>
<td>48.00</td>
<td>33.20</td>
</tr>
<tr>
<td>Nb</td>
<td>0.588</td>
<td>7.00</td>
<td>3.38</td>
<td>1.05</td>
<td>14.30</td>
<td>36.00</td>
</tr>
<tr>
<td>La</td>
<td>0.686</td>
<td>17.88</td>
<td>7.23</td>
<td>1.87</td>
<td>28.90</td>
<td>102.70</td>
</tr>
<tr>
<td>Ce</td>
<td>1.786</td>
<td>34.73</td>
<td>14.34</td>
<td>4.06</td>
<td>59.50</td>
<td>181.50</td>
</tr>
<tr>
<td>Nd</td>
<td>1.327</td>
<td>12.32</td>
<td>5.34</td>
<td>1.83</td>
<td>28.70</td>
<td>68.50</td>
</tr>
<tr>
<td>Sm</td>
<td>0.431</td>
<td>2.07</td>
<td>2.41</td>
<td>0.61</td>
<td>6.19</td>
<td>12.96</td>
</tr>
<tr>
<td>Yb</td>
<td>0.462</td>
<td>0.78</td>
<td>2.60</td>
<td>0.58</td>
<td>5.10</td>
<td>2.39</td>
</tr>
<tr>
<td>V</td>
<td>46.00</td>
<td>230.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$^{87}$Sr/$^{86}$Sr</td>
<td>0.704500</td>
<td>0.701116</td>
<td>0.701116</td>
<td>0.708425</td>
<td>0.716805</td>
<td>0.706618</td>
</tr>
<tr>
<td>$^{143}$Nd/$^{144}$Nd</td>
<td>0.512640</td>
<td>0.512150</td>
<td>0.512150</td>
<td>0.512469</td>
<td>0.511776</td>
<td>0.512518</td>
</tr>
</tbody>
</table>
Conclusions

and primary mantle (PM) are reported in Table 4; (C) source (mixing trends). For comparison we have also added the AFC trend like as that described in Fig. 13. The geochemical composition of sediment melts (SM), sediment fluids (SF) compositions. The compositions of primary mantle (PM), Mixed Mantle Source (MMS), sediment melts (SM) and fluids (SF) are from Ersoy et al. (2010) and reported in Table 4; (D) Identification of mantle components of the Saray volcanic rocks using Nb/U vs. Ce/Yb vs. Sr/Yb diagrams to evaluate the effect of sediment fluids and sediment-derived melts on the mantle wedge (primary mantle; PM)

Identification of mantle components of the Saray volcanic rocks using Nb/U vs. Ce/Yb vs. Sr/Yb diagrams to evaluate the effect of sediment fluids and sediment-derived melts on the mantle wedge (primary mantle; PM)

Modified after Prelević et al. (2012).

Quaternary period (e.g., Didon and Germain, 1976; Dostal and Zebri, 1978). This temporal shift from a lithospheric to an asthenospheric source is correlated to asthenospheric upwelling through a tear on the subducted slab of the Arabian plate (Prelević et al., 2012).

7. Conclusions

The Saray high-K rocks in NW Iran yield Ar–Ar ages of ~11 Ma (Late Miocene) and comprise 1) tephriti, phonolitic, trachytic and leucititic lavas, 2) huge volumes of pyroclastic rocks and 3) trachytic, basanitic and lamproitic dikes cutting the above rock units. These rocks are highly enriched in incompatible trace elements and have highly radiogenic Sr and Pb isotopes with non-radiogenic Nd isotopes. The rocks contain clinopyroxene phenocrysts with complex zoning including Al–Ti rich and/or Al-Ti poor varieties. The geochemical features and modeling results indicate that these rocks were produced by the partial melting of refertilized subcontinental mantle material was probably subcontinental mantle that was isolated from the asthenospheric mantle mixing at the time of initial crust stabilization. Subsequently, the subcontinental mantle was enriched by subduction-related fluids or melts. Finally, the enriched mantle underwent low degrees of melting (probably ~2–10%) resulting in the production of high-K rocks. Some of these melts interacted with the crust during their ascent to the surface, resulting in the incorporation of crustal components.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gr.2013.09.015.

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Fig. 14. (A and B) Th/Nb vs. Ba/Nb and Ce/Yb vs. Sr/Yb diagrams to evaluate the effect of sediment fluids and sediment-derived melts on the mantle wedge (primary mantle; PM)
Fig. 15. Trace element non-modal dynamic melting models (with critical melt porosity of 1.1%) of the Saray lavas for garnet bearing amphibole lherzolites from Ulten Zone (Sample MK5C; Rampone and Morten, 2001) with varying amount of garnet (6, 7 and 10%) (See Table 5 for composition and melt mode). For comparison the melting trend of garnet lherzolite composition of Walter (1998) is shown (See Table 5 for composition and melt mode).

Table 5
The composition and melting mode of the garnet lherzolite (Walter, 1998) and the Ulten Zone (UZP) (Rampone and Morten, 2001) mantle sources.

<table>
<thead>
<tr>
<th>Source mode</th>
<th>Garnet–lherzolite (Walter, 1998)</th>
<th>UZP (sample MK5C) 7% garnet</th>
<th>UZP (sample MK5C) 5% garnet</th>
<th>UZP (sample MK5C) 3% garnet</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine</td>
<td>57.0%</td>
<td>54.0%</td>
<td>54.0%</td>
<td>54.0%</td>
</tr>
<tr>
<td>Cpx</td>
<td>21.0%</td>
<td>21.0%</td>
<td>21.0%</td>
<td>21.0%</td>
</tr>
<tr>
<td>Cpx</td>
<td>16.0%</td>
<td>12.0%</td>
<td>14.0%</td>
<td>15.0%</td>
</tr>
<tr>
<td>Garnet</td>
<td>6.0%</td>
<td>7.0%</td>
<td>5.0%</td>
<td>3.0%</td>
</tr>
<tr>
<td>Amphibole</td>
<td>0.0%</td>
<td>6.0%</td>
<td>6.0%</td>
<td>6.0%</td>
</tr>
</tbody>
</table>

Melt mode

| Olivine      | 8.0%                            | 5.0%                         | 5.0%                         | 5.0%                         |
| Cpx          | −19.0%                          | 5.0%                         | 5.0%                         | 5.0%                         |
| Cpx          | 81.0%                           | 30.0%                        | 30.0%                        | 30.0%                        |
| Garnet       | 30.0%                           | 20.0%                        | 20.0%                        | 20.0%                        |
| Amphibole    | 0.0%                            | 40.0%                        | 40.0%                        | 40.0%                        |

Trace elements composition of the Ulten Zone Peridotites (Eastern Italian Alps)

| Th        | 0.620                          |
| La        | 2.420                          |
| Tb        | 0.050                          |
| Yb        | 0.220                          |

Trace elements composition of garnet lherzolite of Walter (1998)

| Th        | 0.029                          |
| La        | 1.681                          |
| Tb        | 0.069                          |
| Yb        | 0.259                          |

Conticelli, S., Peccerillo, A., 1980. Petrological significance of high-pressure ultramafic xenoliths from ultrapotassic rocks of Central Italy. Lithos 24, 23–51.

References
