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Geochemistry of ultramafic, mafic, and felsic xenoliths from the Gölcük (Isparta, SW Turkey) alkali rocks: genetic relationship with arc magmas

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Abstract
In the Isparta Province (SW Anatolia), the Gölcük volcanism comprises of many medium-to-small volume trachyte, trachyandesite, basaltic trachyandesite, phonolite, lamprophyre dome and dykes, and pyroclastic deposits. The domes and dikes together with the explosive volcanism have been occurred at various time intervals throughout the geological history and contain three groups of xenoliths felsic, mafic, and ultramafic. Felsic xenoliths with syenitic and syeno-dioritic composition have inequigranular textures. Mafic xenoliths consist of inequigranular textured gabbroic and monzodioritic rocks. The third group is ultramafic xenoliths with pyroxenitic, equigranular textures. The Gölcük volcanic rocks have intermediate to basic compositions, having shoshonitic to ultrapotassic characteristics. Major element modeling shows that fractional crystallization is the main petrogenetic process in the evolution of the magma. They have high large ion lithophile (LIL) element (Th, K, Sr, Ba) concentrations but are low in high field strength (HFS) element (Nb, Ta, P, Ti, Zr). The extreme depletion in the elements indicates that there is a partially modified mantle source with a subduction. All rocks are enriched in light rare earth elements (LREE) and do not show any Eu anomaly. For this reason, it is generally thought that all rocks (xenolith types and host rocks) were formed from similar magma. The magma rise to shallow magma chamber formed of fractional crystallization from a deep magma chamber before eruption.

Keywords Gölcük · Volcanism · Xenolith · Isparta

Introduction

The Gölcük (Isparta) volcanism has three different stages. These are (i) extrusive volcanism consisting of trachyte, trachyandesite, basaltic trachyandesite, phonolite, tephrite, and lamprophyres; (ii) explosive volcanism composed of ignimbrite, unconsolidated tuff, agglomerate, and pumice; and (iii) extrusive volcanism consisting of trachyte and trachyandesite. The trachyte, trachyandesites, and their pyroclastics contain ultramafic, mafic, and felsic xenoliths having phaneritic texture.

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Lacroix (1893) proposed an enclave term to describe rock fragments in magmatic rocks. Lacroix enclaves are described in two groups which are related to the parent rock in terms of both genetic and mineralogical (1), and their origin, mineralogical, and chemical composition (2). Sollas (1894) used the term "xenolith" for the incoming rock fragments and "xenocryst" for the separated crystals that appeared in the magmas. Then, instead of homologous enclaves, "cognate enclaves" term, and instead of heterogeneous enclaves, "foreign enclaves or xenoliths" terms were used (Nitoi et al. 2002).

In many petrology literatures, mafic microgranular enclaves (MMEs) (Fershtater and Borodina 1977; Vernon 1984, 1990; Frost and Mahood 1987; Bedard 1990; Dodge and Kistler 1990; Dorais et al. 1990; Poli and Tommasini 1991; Srogi and Lutz 1990; Castro et al. 1991; Waight et al. 2001; Barbarin and Didier 1992; D'Lemos 1996; Nardi and Lima 2000; Silva et al. 2000; Janoušek et al. 2004; Barbarin 2005; Rajaieh et al. 2010) are defined as materials of relatively fine grain size, generally elliptical, with typical magmatic microstructure, and mafic mineral groups. They are more prevalent in granitoid plutons and their equivalent surface rocks

(Cantagrel et al. 1984; Bacon 1986; Didier and Barbarin 1991; Rajaieh et al. 2010). Many hypotheses have been proposed for the origin of MMEs. The hybridization model has been taken into account by many petrologists (e.g., Vernon 1983, 1990; Cantagrel et al. 1984; Didier 1984; Ayrton 1988; Vernon et al. 1988; Dorais et al. 1990; D'Lemos 1996; Rajaieh et al. 2010) and applied to the study of granites in the worldwide. These scientists have argued that MMEs are mafic magma bubbles that are mixed or partially mixed with felsic solutions of crust, possibly mantle-produced and coming from outside. Contrary to previous ideas, some researchers suggested the restite model (White and Chappell 1977; Chappell et al. 1987; White et al. 1999). Based on this model, the mafic inclusions are restite (residual source material) and therefore have derived from the source of partial melting. They represent parts of that source that have remained coherent during melting and inclusion (Rajaieh et al. 2010).

The Gölcük volcanism, which formed during the Plio-Pleistocene timespan, provides valuable information on petrological, geochemical, and geochronological history (Platevoet et al. 2008). Many literatures mainly focus on the petrology and evolution of volcanism, and regional tectonic setting (e.g., Lefèvre et al. 1983; Alıcı et al. 1998; Coban 2005; Kumral et al. 2006, 2007; Platevoet et al. 2008; Elitok et al. 2010; Platevoet et al. 2014; Schmitt et al. 2014). Studies on xenoliths of this volcanism and their relations with their host rocks and their origins are rare (Sen et al. 2008; Platevoet et al. 2014). In the worldwide, many articles related to the enclaves which are usually found in granitoid intrusions have been discussed. However, the xenoliths in the rocks such as trachyte, trachyandesite, trachybasalt, and pyroclastic have been much less noticeable.

It is thought that xenoliths are the source of important information on the processes that define the evolution of the magmas in the study area. Taking into account their features and relationships with the host rocks, the xenoliths provide valuable information about magmatic processes, such as fractional crystallization and assimilation, and geotectonic environments. In this study, mineralogical, petrographic, and geochemical features of xenoliths found in domes, dykes, and pyroclastics are described. The aim is to reveal a general description of xenoliths and to discuss their origin and evolution.

Regional geology

Autochthonous units

Kayıköy formation

The formation with flysch characteristics was named by Karaman et al. (1988). The unit is composed of sandstone, conglomerate, siltstone, claystone, limestone, clayey limestone, and mudstones. The age of the unit is middle-late Eocene (Karaman 2000; Görmüş and Özkul 1995). It is overlaid with Gökçebağ melange and Akdağ limestones belonging to Lycian nappes (Fig. 1). There are lavas and pyroclastics of the Gölcük volcanism on the unit.

Gölcük Volcanics

These volcanics were identified as Gölcük Volcanics by Poisson (1977). Volcanism is a centre of maar-type volcanic eruption (Kazancı and Karaman 1988; Nemec et al. 1998). Volcanic rocks consist of domes and dykes shaped lavas and pyroclastic rocks, such as volcanic block-bomb, lapilli, pumice, ignimbrite, agglomerate, and unconsolidated tuff. Both the host rocks and the pyroclastics contain xenolithes with difference properties. Different types of xenoliths can be distinguished by their grain size, texture, structure, mineral content and abundance of phenocrysts, composition, and contacts with host rocks. Using origin and textural criteria, xenoliths were grouped into three types: (i) felsic xenoliths coarse-to-medium, rarely fine-grained, phaneritic textured, syenitic, and syeno-dioritic compositions of the same origin as the host magmas; (ii) mafic xenoliths medium-to-fine, rarely coarse grained, holocrystalline, monzodioritic, and gabbroic; and (iii) ultramafic xenoliths of medium-to-fine, rarely coarse-crystals, phaneritic textured, pyroxenitic. Xenoliths are angular, sometimes rounded and lenticular in shape, on the host rocks ranging in size from a few centimeters to 15–20 cm, in pyroclastics 45–50 cm, seldom larger.

Felsic xenoliths are seen as aggregates in pyroclastics and volcanic rocks. These are coarse-medium grained, light colored, in size ranging from a few centimeters to 10–15 cm (Fig. 2a). Mafic and ultramafic species are medium-to-fine, rarely coarse-grained, dark colored, in size ranging from a few centimeters to 45–50 cm (Fig. 2b, c). Their definition as xenoliths is supposed to be that the contacts with the host rocks are generally sharp and that no reaction zone is visible in the transition regions.

The age of the volcanics has been determined in many studies (Lefèvre et al. 1983; Platevoet et al. 2008, 2014; Mouillard 2011; Schmitt et al. 2014) and has been the subject of some studies (Nemec et al. 1998; Alıcı et al. 1998; Kumral et al. 2007; Elitok et al. 2010). Gölcük volcano is a multi-cycle eruptive center with a prolonged history of explosive and effusive eruptions throughout the Pleistocene (Alıcı et al. 1998; Platevoet et al. 2008; Elitok et al. 2010; Schmitt et al. 2014). Previous $^{40}K/^{40}Ar$ dating was performed by Lefèvre et al. (1983) on lavas showed that the first lava protrusions occurred between 4.7 and 4 Ma. According to Coban (2005), the ages were found from all volcanic rocks (e.g., lamprophyric dykes, trachyandesite and trachyte, basaltic-trachyandesite and trachybasalt, pyroclastics, and phonolithic dykes) 6.21 + 0.3–0.35 + 0.1 Ma.

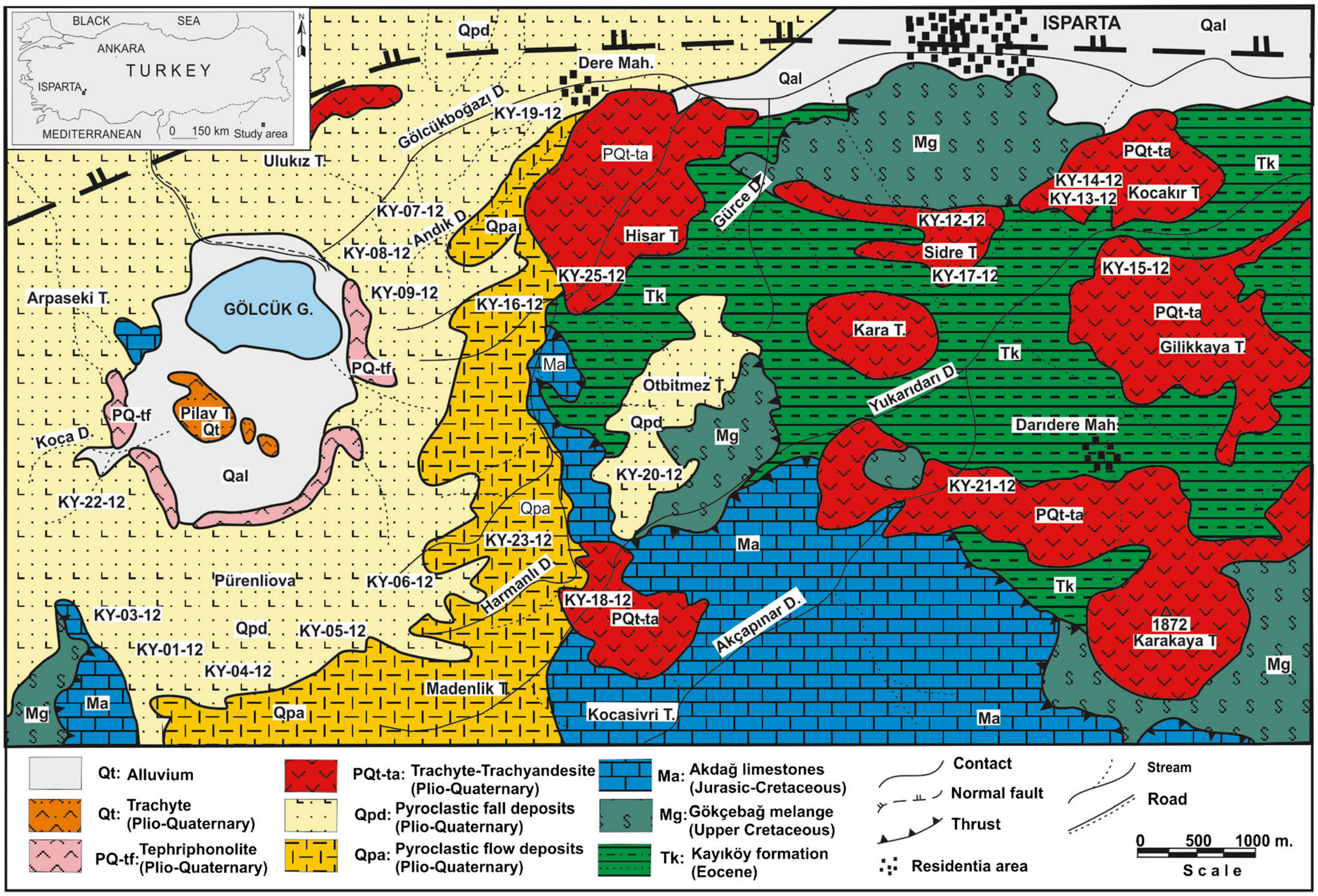

Fig. 1 Location and geological map of the investigated area (modified from Elitok et al. 2008)

Platevoet et al. (2008, 2014) stated that the volcanism of Gölcük was active for three phases and determined the ages of each phases with $^{40}K/^{40}Ar$ and $^{40}Ar/^{39}Ar$ methods. According to this, it is stated that the first period trachyandesitic rocks were formed in 2.77 ± 0.06 Ma and the pyroclastic flow products were between 440 ± 12 ka and 148 ± 21 ka. Tefriphonolite

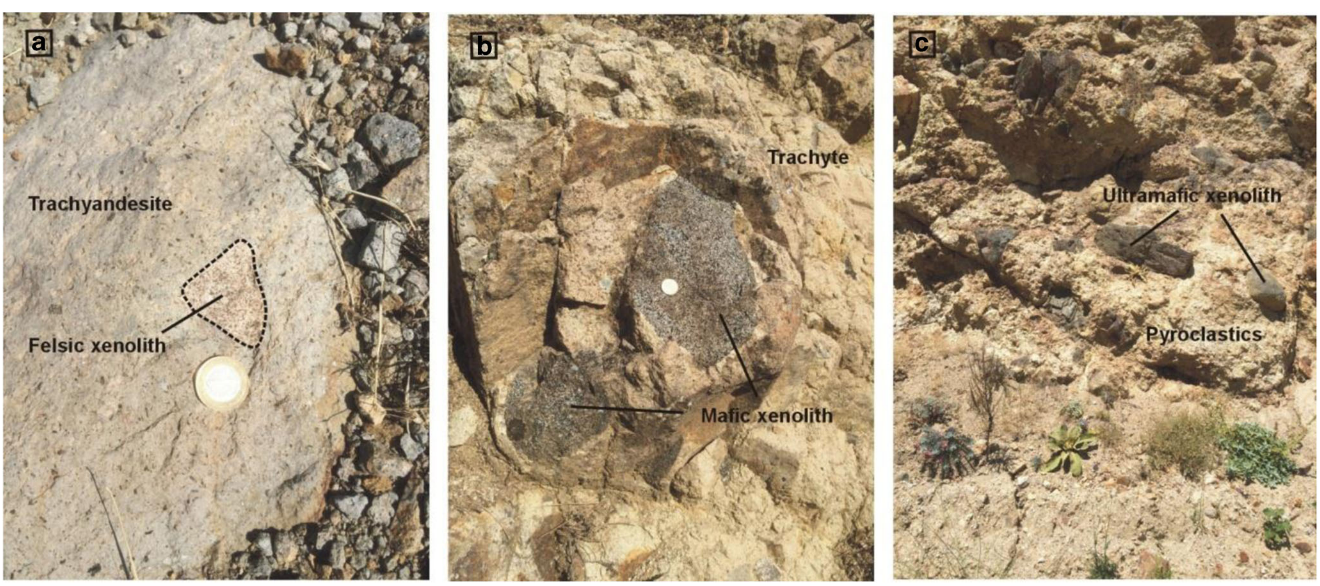

Fig. 2 Macroscopic photos of representative xenoliths from Gölcük area (Isparta). **a** Felsic xenolith with angular, medium-coarse crystal in the host trachyandesites. **b** Mafic xenoliths with sharp and angular contact in the host trachyte. **c** The ultramafic xenolithes are rounded to elliptical in shape in the pyroclastics

dome and dykes belonging to the second period aged between 115 ± 3 ka and 62 ± 2 ka. It is determined that the pyroclastic rocks of the third period were 72.7 ± 4.7 ka and the last settled trachyte dome age was 24 ± 2 ka. The explosive activity started about 2180 ± 44 ka ago (multimodal age by single feldspar $^{40}Ar/^{39}Ar$ method, Mouillard 2011) by huge pyroclastic flows emplacing within the lacustrines of the Burdur lake. Schmitt et al. (2014) had obtained (U-Th)/He zircon eruption ages from two different trachytic lava domes in the caldera to determine when the Gölcük volcanism was the last active. According to this, it is explained that volcanism was last active 14.1 ± 0.5 and 12.9 ± 0.4 ka and that it has been sleeping since then.

Allochthonous units

Gökçebağ Melange

This unit named as Gökçebağ Melange by Sarıız (1985) is an ophiolitic complex consisting of serpentinite, gabbro, diabase, chert, radiolarite, and limestone blocks observed in light-dark green and reddish colors. It overlies overthrustly the Kayıköy Formation. The unit is cut by lavas of Gölcük volcanism and overlain unconformably by pyroclastics.

Akdağ limestone

Carbonaceous rocks overthrust on the Gökçebağ Melange are named by Karaman (1990). Akdağ Limestones and Gökçebağ Melange, known as Likya nappes, are located overthrust with the compressional tectonic that occur in the middle Miocene (Poisson 1984). The unit is cut by lavas of the Gölcük volcanism and covered unconformably by pyroclastics.

Petrography

Petrographical and geochemical studies suggest that there are three types of xenoliths in the Gölcük volcanism. These xenoliths vary in size, color, texture, and composition, and distributed randomly in lava and pyroclastics. They are seen rarely in trachyte, trachyandesite, and basaltic trachyandesites, and are more common in lately occured pyroclastics. These were preserved in magma at different stages of assimilation. Xenoliths belong to three petrographic and geochemical groups: felsic syenites and syeno-diorites, mafic gabbros, and ultramafic pyroxenites. Most xenoliths consist of assemblages of a few mineral phases (such as clinopyroxene, plagioclase, mica) and are similar to those found in their host rocks. Varieties of xenoliths are indicated in Platevoet et al. (2014) and Sen et al. (2008).

Host rocks

Trachyte, trachyandesite, and basaltic trachyandesites

These rocks are the most common rocks of the Gölcük volcanism. They may occur as domes and dykes with country rocks. Their textures are porphyritic and microlitic porphyritic (Fig. 3a). Host rocks of the Gölcük volcanism are composed of alkali feldspar (sanidine) (35–50%), plagioclase (30–50%), clinopyroxene (20–25%), mica (10–15%), and amphibole (10–15%). Apatite and titanite are accessory minerals.

Alkali feldspar (sanidine) crystals occur both as large phenocrysts and microlites in the groundmass. They commonly have anhedral to subhedral forms and are rounded due to corrosion. Sanidine crystals include plagioclase, clinopyroxene, amphibole, mica, apatite, and titanite inclusions, presumably resulting from rapid crystallization. Plagioclase grains occur both as phenocrysts and as microliths in the groundmass. These form with subhedral to euhedral prisms and polysynthetic twin, and some crystals are affected by calcite, epidote, and serizite. Clinopyroxene crystals are especially seen in trachyandesite and basaltic trachyandesites. They are in euhedral and subhedral forms and in fine-to-medium grain sizes. They are medium green to colorless in thin section, and some phenocrysts have a colorless core, light green mantle, and a darker green rim. Abundant amphiboles occur with euhedral to subhedral texture, and they have in medium-small grain sizes. Amphiboles may show resorption feature and opaque reaction rims. They also contain small-sized plagioclase inclusions that have been trapped and prevented to further growing. Micas are abundant and occur as large elongate plate shape, hexagonal flakes. They are usually corroded and may show opaque reaction products near the margin of the crystals. Some mica have a pale yellowish core and dark brown rim. Apatite is a common accessory and may be present both as long prismatic and blunt stubby prisms.

Xenoliths

Ultramafic xenoliths

Ultramafic xenoliths are pyroxenitic compositions. They are composed of clinopyroxene (55–60%) and mica (phlogopite) (35–40%). Small size intergranular apatites are accessory mineral (Fig. 3b). They usually have equigranular texture. Clinopyroxenes form as euhedral and subhedral. They are pleochroic from pale green to colorless.

Mafic xenoliths

Mafic xenoliths are monzodioritic and gabbroic. They display inequigranular texture as grains which range in size from 1.0 to 2.0 mm. Their mineralogical compositions make up mafic

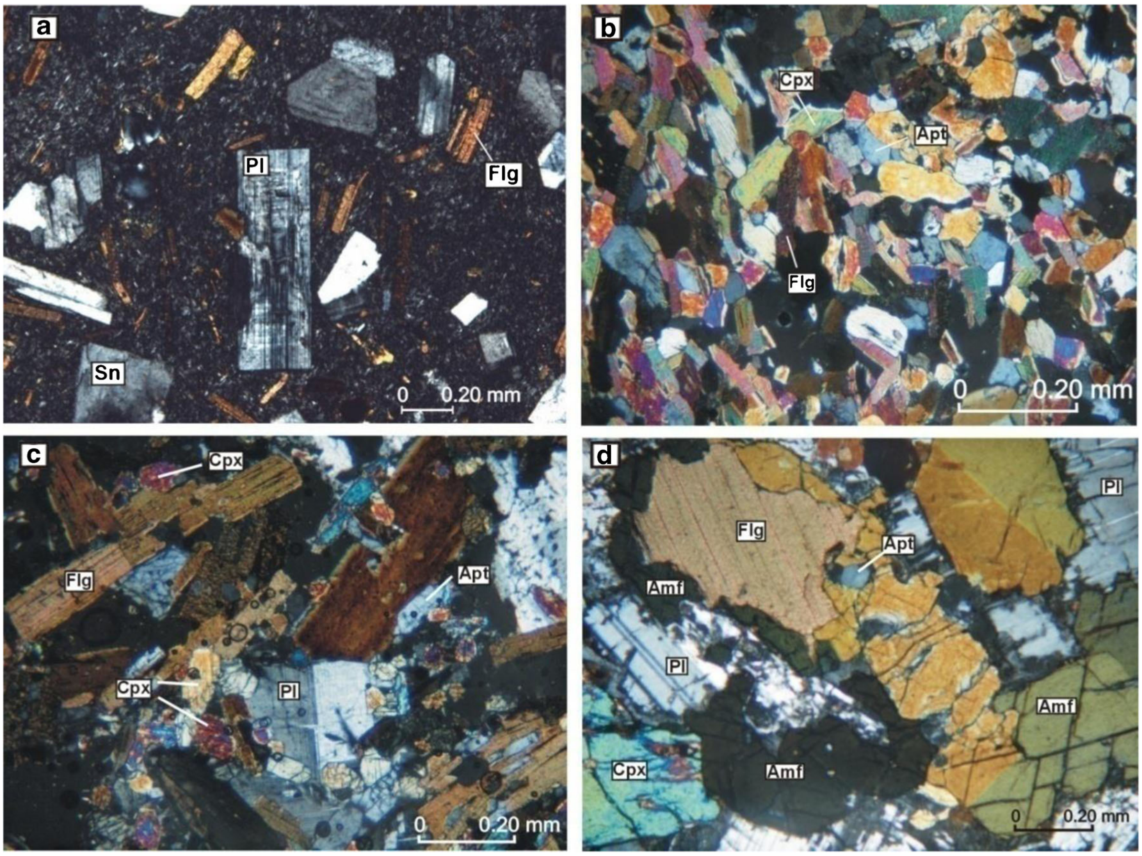

Fig. 3 Photomicrographs of the parent rock and xenoliths. **a** Trachyandesite. Sanidine commonly have anhedral to subhedral forms. Plagioclase grains occur both as phenocrysts and as microliths in the groundmass. These forms with subhedral to euhedral prisms and polysynthetic twin. Mica occurs as large elongate plate shape, hexagonal flakes. **b** Ultramafic xenoliths. Clinopyroxenes occur as subhedral to euhedral, short prismatic, and tabular crystals. Some crystals have simple twins. Phlogopites are usually euhedral and subhedral, long and tabular-shape. Plagioclase phenocrysts show euhedral-subhedral, tabular shapes, polysynthetic twin, and oscillatory zonation. **c** Mafic xenoliths have usually equigranular texture. Clinopyroxenes forms as euhedral and subhedral. They is pleochroic from pale green to colorless. Mica forms euhedral, subhedral, and occurs as large elongate plate shape, hexagonal flakes. Apatite forms euhedral and hexagonal. **d** Felsic xenoliths are medium to coarse grained and with inequigranular textures. K-feldspar forms anhedral, rarely subhedral crystals of orthoclase. Plagioclase mainly forms subhedral to anhedral, prismatic, and lath-shaped crystals. Amphibole forms as subhedral to euhedral large prismatic crystals. Mica (phlogopite) minerals occur as medium and fine grain phenocryst. Some phlogopites is surrounded by hornblende reaction rim. Clinopyroxene phenocrysts are euhedral to subhedral and often show oscillatory zoning. (Pl plagioclase, Cpx clinopyroxene, Flg phlogopite, Sn Sanidine, Amf amphibole, Apt apatite)

minerals such as clinopyroxene and phlogopite (80–90% of rock volume), and plagioclases (15–20% of rock volume) (Fig. 3c). Apatite and opaque are accessory minerals.

The clinopyroxenes occur as subhedral to euhedral, short prismatic, and tabular crystals. Some crystals have simple twins. Phlogopites are usually euhedral and subhedral, long and tabular-shape. Plagioclase phenocrysts show euhedral-subhedral, tabular shapes, polysynthetic twin, and oscillatory zonation. They are not much affected by alteration. Apatite and titanite are accessory microphenocrysts. Apatite occurs as inclusion within clinopyroxene, plagioclase, and phlogopite.

Felsic xenoliths

The felsic xenoliths (syenite and syeno-diorite) are plutonic products exhibiting petrographic features similar to the host rocks. They were formed before the volcano and consist mainly of K-feldspar (orthoclase), plagioclase, clinopyroxene, amphibole, and mica (phlogopite) (Fig. 3d) in different proportions. Apatite and opaque oxides are accessory minerals. They are medium- to coarse-grained and with inequigranular textures. K-feldspar forms anhedral, rarely subhedral crystals of orthoclase. Large pheno-crysts (greater than 2 mm) are heterogeneously distributed within the various xenoliths. Alteration to clay minerals is more common in large K-feldspar (orthoclase) crystals than in plagioclase. Plagioclase mainly forms subhedral to anhedral, prismatic, and lath-shaped crystals. They mainly show oscillatory zoning, polysynthetic twin, and prismatic-cellular growth. Some large plagioclase crystals may contain clinopyroxene, phlogopite, amphibole, and apatite showing poikilitic texture. Amphibole forms as subhedral to euhedral large prismatic crystals. They are commonly zoned and twinned. Mica (phlogopite) minerals occur as medium and fine grain phenocryst. They display distinct pleochroism with light yellow to yellowish-brown color. Some phlogopites are surrounded by hornblende reaction rim. Clinopyroxene phenocrysts are euhedral to subhedral and often show oscillatory zoning. Some crystals have a colorless core and light green mantle, and there are simple twins. Apatites have thin-medium size and euhedral shape. Some of the crystals are corroded and rounded.

Analytical methods

The major- and trace-element composition of 21 xenolith samples was analyzed by Inductively Coupled Plasma Emission Spectroscopy (ICP-ES) at Acme Analytical Laboratories Ltd., Canada. The detection limits of ICP-ES for major, trace, and rare-earth elements (REE) are typically in the range of 0.04–0.01 wt%, 1–0.5 ppm, and 0.5–0.01 ppm, respectively.

All rock chemistry

Major elements

Major, trace, and rare earth elements (REE) are analyzed from ultramafic, mafic, and felsic xenoliths (Table 1). SiO_2 content ranges from 37.5 to 49.7% in mafic and ultramafic xenoliths, from 50.8 to 61.9% in felsic xenoliths, and from 47.8 to 66.3% in host rocks. Major and minor element variation diagrams for host rock and xenoliths are given in this study. In TAS diagram (Fig. 4), the ultramafic xenoliths plot in the fields of tephrite-basanite and foidite, mafic xenoliths in basalt and trachy-basalt, and felsic xenoliths in trachyte and trachyandesite compositions. On the other hand, the host rocks differentiate mostly into the basaltic trachyandesite, trachyandesite, and trachyte with alkalic affinities. The K_2O vs. SiO_2 diagram (Peccerillo and Taylor 1976) shows that all samples are shoshonitic and ultrapotasic compositions (Fig. 5).

The following values were determined: 0.10–4.85% CaO, 1.41–9.12% MgO in host rocks; 8.37–25.93% MgO, 11.09–17.33% CaO (except to KY-05 = 1.21%) in ultramafic xenoliths; 7.87–16.96% MgO, 11.09–17.33% CaO in mafic xenoliths; and 1.34–3.15% MgO, 2.13–8.10% CaO in felsic xenoliths. The ultramafic and mafic xenoliths that plot upper the host rocks and felsic xenoliths trend are enriched with phlogopite and clinopyroxene (Fig. 6). Differences between host rocks and felsic xenoliths with dark xenoliths are related to the relative abundance of mafic minerals in dark xenoliths. The host rocks and felsic xenoliths have higher SiO_2, K_2O, and Na_2O; and lower CaO, Fe_2O_3, MgO, TiO_2, and P_2O_5 than that ultramafic and mafic xenoliths. Harker variation diagrams (Fig. 7) show decreasing trend for Fe_2O_3, CaO, MgO, TiO_2, MnO, and P_2O_5 with increasing silica for all host rocks and xenoliths. On the contrary, Al_2O_3 values show increasing content with increasing silica from ultramafic and mafic xenoliths to host rocks and felsic xenoliths, whereas the felsic xenoliths show flat correlation with the host rocks. Na_2O and K_2O values show a positive correlation with increasing silica in the host rocks and xenoliths. The Mg numbers of ultramafic, mafic, and felsic xenoliths and host rocks display values between 63 and 91, 63 and 86, 35 and 67, and 23 and 61, respectively.

Trace elements

Compared to the felsic xenoliths and host rocks, ultramafic and mafic xenoliths generally show higher abundances in compatible elements such as Cr, Ni, and Sc (Fig. 8). This suggests the fractionation of olivine and clinopyroxene. Sr show increasing in the ultramafic and mafic xenoliths, whereas decreasing in the host rock and felsic xenoliths. The high abundant of Zr in all rock types is as a result of open-system fractional crystallization, in which the fractionation of amphibole and clinopyroxene fractionation significantly increase Zr content. The Ba shows a flat correlation.

The primitive mantle-normalized multielement variation diagrams of the host rocks and the xenoliths show positive anomalies for large ion lithophile elements (LILE) (e.g., Ba, Sr, K, Th), and marked negative anomalies for high-field strength elements (HFSE) (e.g., Nb, Ta, Ti, and P).

These may include negative Nb and Ta spikes in the spider diagram, which may suggest either crustal contamination (Cox and Hawkesworth 1985; Peng et al. 1994) or arc volcanism enriched by subduction processes and slab-derived fluids (Fitton et al. 1988; Saunders et al. 1988). The depletion of Ti and P indicates substantial fractionation of Fe-Ti oxides and apatite, respectively, from the melt. However, the enrichment patterns are similar in the host rocks and xenoliths (Fig. 9).

The chondrite-normalized rare earth elements (REE) patterns for the host rocks, and xenoliths are presented in Fig. 10. The diagram shows that all the rocks are enriched. Ultramafic xenoliths have light rare earth elements (LREE) abundances La_N = 354–875, except for sample KY-05 (La_N = 30,071). Mafic and felsic xenoliths, and host rocks have varying abundances of LREE (La_N = 84–680, La_N = 304–1165, and La_N = 232–865, respectively). Similarly, ultramafic xenoliths have heavy rare earth element (HREE) abundances Yb_N = 6.5–12.5, except for sample KY-05 (Yb_N = 108.0). Mafic and felsic xenoliths, and host rocks have abundances of HREE (Yb_N = 1.7–12.7, Yb_N = 7.0–12.9, and Yb_N = 3.6–11.4, respectively).

The ultramafic, mafic, and felsic xenoliths and host rocks represent wide-range variation ratios [$(La/Yb)_N$ = 43.7–97.4, 28.5–72.7, 33.7–90.6, and 32.9–85.2, respectively] in LREE/HREE. This wide range of values is possibly present

Table 1 Whole-rock major (wt%) and trace (ppm) element analyses of representative samples of the host rocks, ultramafic, mafic, and felsic xenoliths from the Gölcük volcanics. Mg# = atomic 100 × (Mg/Mg + Fe^{2+}), assuming Fe_2O_3/FeO = 0.1. Chemical compositions of host rocks are from Elitok et al. (2010)

Sample	Gl-40	Gl-48	Gl-55	Gl-70	Gl-81	Gl-96	Gl-109
Host rock	Basaltic-trachyandesite						
SiO_2 (wt%)	50.18	51.29	51.92	47.81	50.60	51.19	51.12
Al_2O_3	16.34	16.09	16.51	17.14	16.15	15.98	17.43
Fe_2O_3	6.17	5.59	6.01	7.14	5.79	6.50	6.27
MgO	4.33	3.57	4.52	4.85	3.44	2.87	3.59
CaO	6.46	9.12	8.25	7.76	9.06	7.51	8.18
Na_2O	1.84	2.22	2.74	2.25	3.24	1.63	4.56
K_2O	5.06	5.74	4.82	3.64	4.56	4.44	3.30
TiO_2	1.41	0.82	0.78	1.47	0.65	0.97	0.71
P_2O_5	1.05	0.69	0.64	1.11	0.61	0.87	0.70
MnO	0.07	0.11	0.11	0.09	0.12	0.11	0.12
LOI	6.5	4.0	2.8	6.0	5.1	6.5	3.0
Total	99.42	99.25	99.11	99.28	99.32	98.57	98.98
Mg#	58.0	56.0	60.0	57.7	54.0	46.9	53.0
Sr (ppm)	3143.7	4362.8	5390.0	3067.6	3530.8	6535.9	5678.5
Ba	2418.2	2951.0	2835.0	2910.9	3495.6	5351.6	3586.2
Rb	118.5	149.6	102.7	42.4	107.2	28.8	130.9
Cs	1.5	3.8	1.7	0.7	3.9	7.5	3.4
Ga	18.9	19.3	19.1	21.2	21.2	23.1	22.2
Hf	8.3	10.1	7.8	9.9	9.3	20.4	9.3
Ni	54	12	19	58	10	8	13
Sc	13	11	14	15	13	9	12
Cr	34	14	68	34	34	14	14
Zr	360.2	427.5	336.9	406.6	386.1	817.3	415.6
Y	25.7	30.4	25.7	26.2	28.3	40.5	29.3
Nb	55.8	47.4	45.3	61.8	50.7	90.9	51.8
Ta	3.1	2.7	2.9	3.5	3.1	5.0	3.4
Th	31.2	48.4	43.6	33.8	55.2	66.4	66.6
U	4.7	12.7	8.8	5.3	13.1	4.7	14.0
Pb	6.0	6.1	3.7	8.1	11.3	18.7	1.7
La	165.3	223.6	196.1	174.9	189.3	317.4	297.6
Ce	312.9	405.5	354.3	325.5	345.2	610.5	550.7
Pr	32.46	42.29	36.17	33.62	33.89	62.36	54.21
Nd	114.3	145.3	118.5	113.7	115.0	215.4	188.7
Sm	16.2	19.3	16.4	16.5	16.4	30.4	23.5
Eu	4.17	4.83	4.15	4.48	3.98	7.69	5.79
Gd	8.78	9.55	8.41	9.35	8.50	15.13	10.08
Tb	1.18	1.25	1.22	1.13	1.16	2.11	1.25
Dy	5.29	5.81	5.25	5.50	5.38	8.07	5.61
Ho	0.90	0.93	0.84	0.88	0.87	1.18	0.93
Er	1.99	2.50	2.15	2.21	2.36	3.04	2.26
Tm	0.31	0.36	0.35	0.31	0.32	0.44	0.36
Yb	1.65	2.24	2.09	2.11	2.17	2.80	2.36
Lu	0.25	0.33	0.31	0.26	0.34	0.44	0.34
$(La/Sm)_N$	6.42	7.29	7.53	6.67	7.27	6.57	7.97
$(Gd/Yb)_N$	4.31	3.46	3.26	3.59	3.17	4.38	3.46
Eu/Eu*	1.07	1.09	1.08	1.10	1.03	1.09	1.15
La_N	450	609	534	477	516	865	811
Yb_N	6.7	9.0	8.4	8.5	8.8	11.3	9.5
$(La/Yb)_N$	67.7	67.5	63.4	56.0	58.9	76.6	85.2
Ba/Nb	43.34	62.26	62.58	47.10	68.95	58.87	69.23
La/Nb	2.96	4.72	4.33	2.83	3.73	3.49	5.75
Rb/Nb	2.1	3.2	2.3	0.7	2.1	0.3	2.5
Th/Ta	10.06	17.93	15.03	9.66	17.81	13.28	19.59
Th/Nb	0.56	1.02	0.96	0.55	1.09	0.73	1.29
Rb/Sr	0.038	0.034	0.019	0.014	0.030	0.004	0.023

of residual garnets in the source. Since garnets have very low partition coefficients ($D << 1$) for LREE and significantly greater compatibility for HREE (Wilson 1989). Buffering of HREE and increasing La/Yb fractionation with increasing overall enrichment in incompatible trace elements in Gölcük host rocks and xenoliths suggest a systematic variation in the

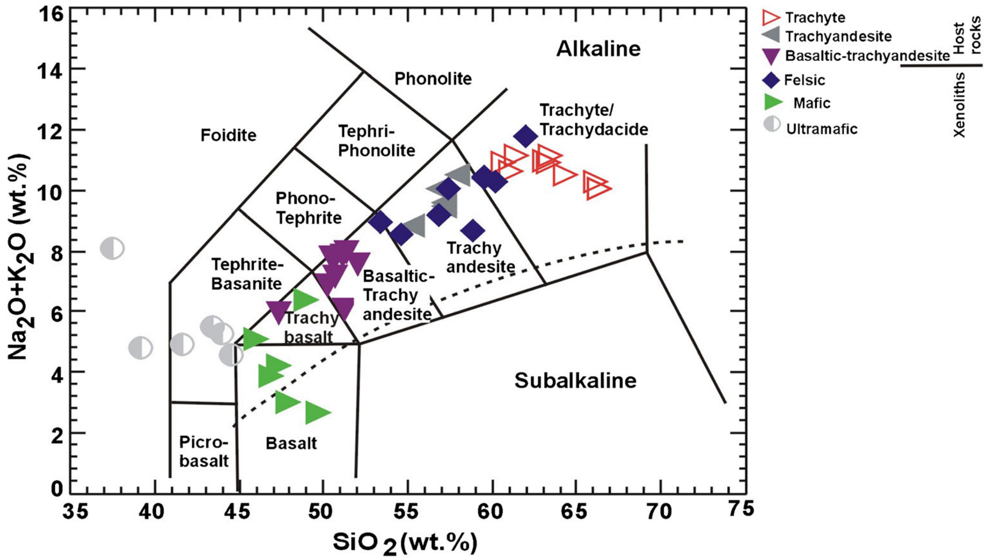

Fig. 4 TAS diagram of host rocks and xenoliths from the Gölcük (Le Maitre et al. 1989)

degree of partial melting of a garnet-bearing source during their petrogenesis (Yılmaz 2010).

The slight positive Eu anomalies in the pattern (Eu/Eu* = 0.82–99.0, 0.88–1.02, 0.86–1.03, and 0.95–1.40 for the ultramafic, mafic, and felsic xenolith and host rocks, respectively) are the result of plagioclase accumulations that contributed to light REE abundances and Eu, leading to the dilution of the HREE (Hassanipak et al. 1996).

Discussion

On the basis of numerous studies on xenoliths worldwide, they are associated with magmatism formed in a volcanic arc setting. In this study, the xenoliths in the studied area are extrusive and explosive volcanism products. In the literature, the Gölcük volcanism was reported as a volcanic arc (Alıcı et al. 1998; Aldanmaz et al. 2006; Coban and Flower 2006; Dilek and Altunkaynak 2007, 2008; Elitok et al. 2010). The characteristics of volcanism and xenoliths are discussed below.

Field relationship

The Gölcük volcanism consists of extrusive rocks including trachyte, trachyandesite, basaltic trachyandesite, and pyroclastic rocks comprising ignimbrite, ash tuff, agglomerate, and lahar. Ash tuff and trachyandesite rocks are the

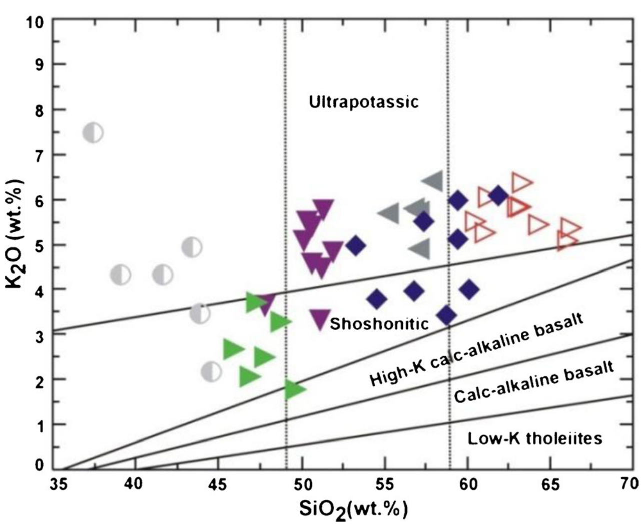

Fig. 5 K_2O versus SiO_2 diagram (Peccerillo and Taylor 1976) of host rocks and xenoliths from the Gölcük. Symbols are as in Fig. 4

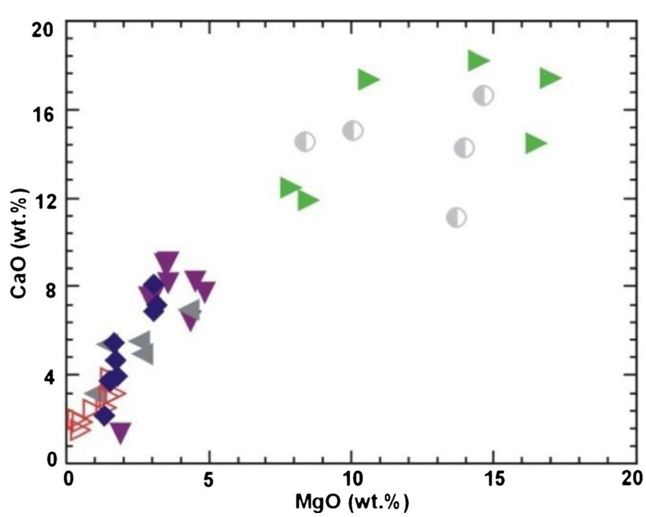

Fig. 6 MgO versus CaO diagram for the host rocks and xenoliths of the Gölcük. Symbols are as in Fig. 4

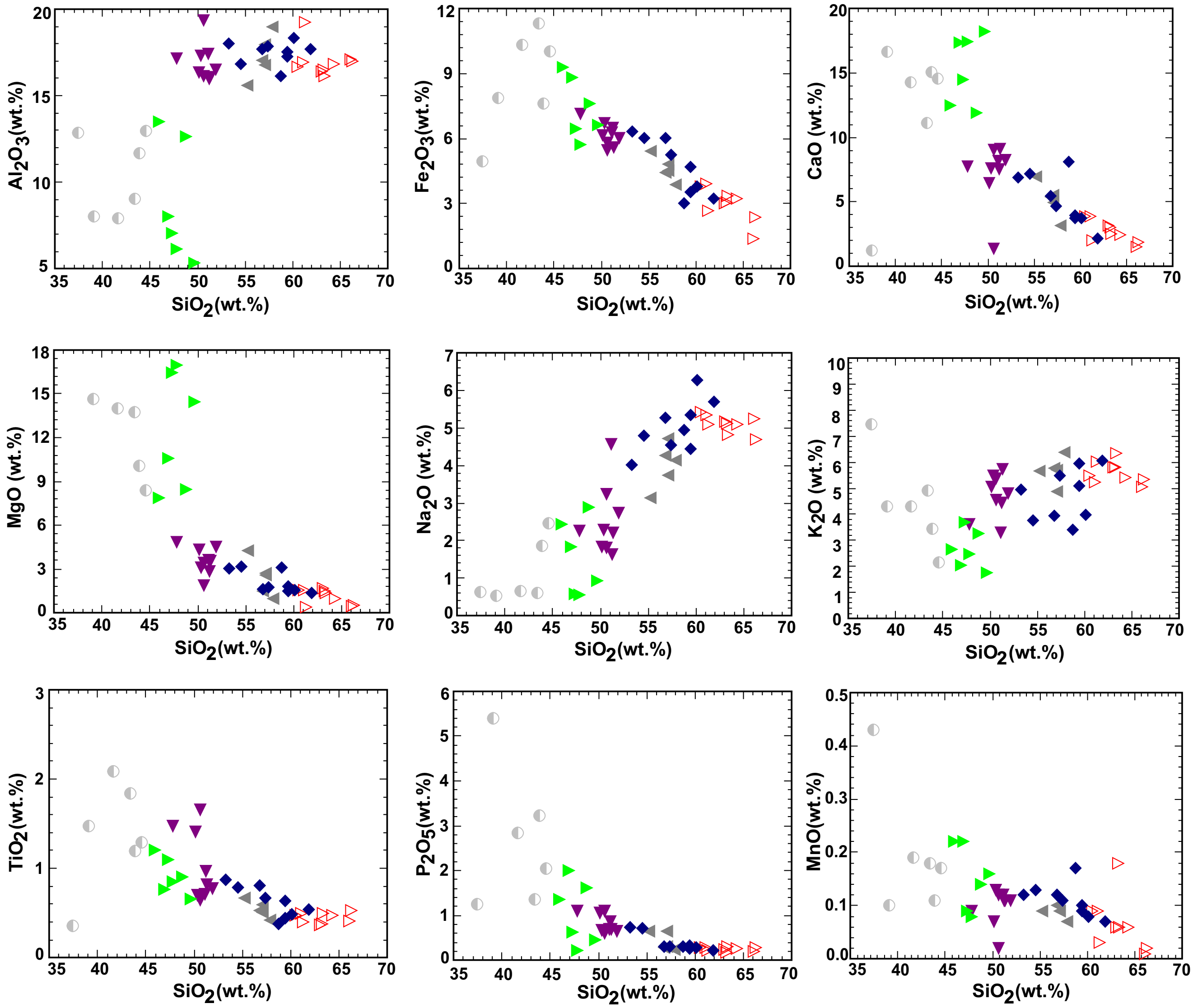

Fig. 7 Major oxide Harker variation diagrams from Gölcük host rocks and xenoliths. Symbols are as in Fig. 4

most common volcanic rocks, whereas phonolite, lamprophyre, and tephriphonolite rocks are less common. Xenoliths have different sizes and shapes, and they are the most common in the pyroclastics, whereas less common in the host rocks. Mafic and ultramafic xenoliths are rounded to elliptical shape, medium-to-fine, rarely coarse-grained, ranging from a few centimeters to 45–50 cm in sizes. Felsic ones are coarse-medium grained, light colored, ranging from a few centimeters to 10–15 cm in sizes. Relationships to the host rocks and pyroclastics of xenoliths include the following: (1) The margins with the host rocks are very sharp and sometimes angular, (2) they are found as nodules and occur only as single xenoliths, and (3) they are sparse and randomly distributed.

Petrographic features

Mineralogical properties

Details of the petrography from the studied xenoliths and host rocks are given in earlier papers (Platevoet et al. 2014; Elitok et al. 2010, respectively), and so only a brief summary is given here. The ultramafic xenoliths are composed mainly of clinopyroxene (diopside) and mica (phlogopite), while mafic ones are formed from plagioclase (oligoclase, andesin) together with these dark-colored minerals. They contain apatite and opaque minerals as accessory. Felsic xenoliths composed mainly of K-feldspar (orthoclase), plagioclase (albite, oligoclase, andesine), clinopyroxene (diopside), mica (phlogopite),

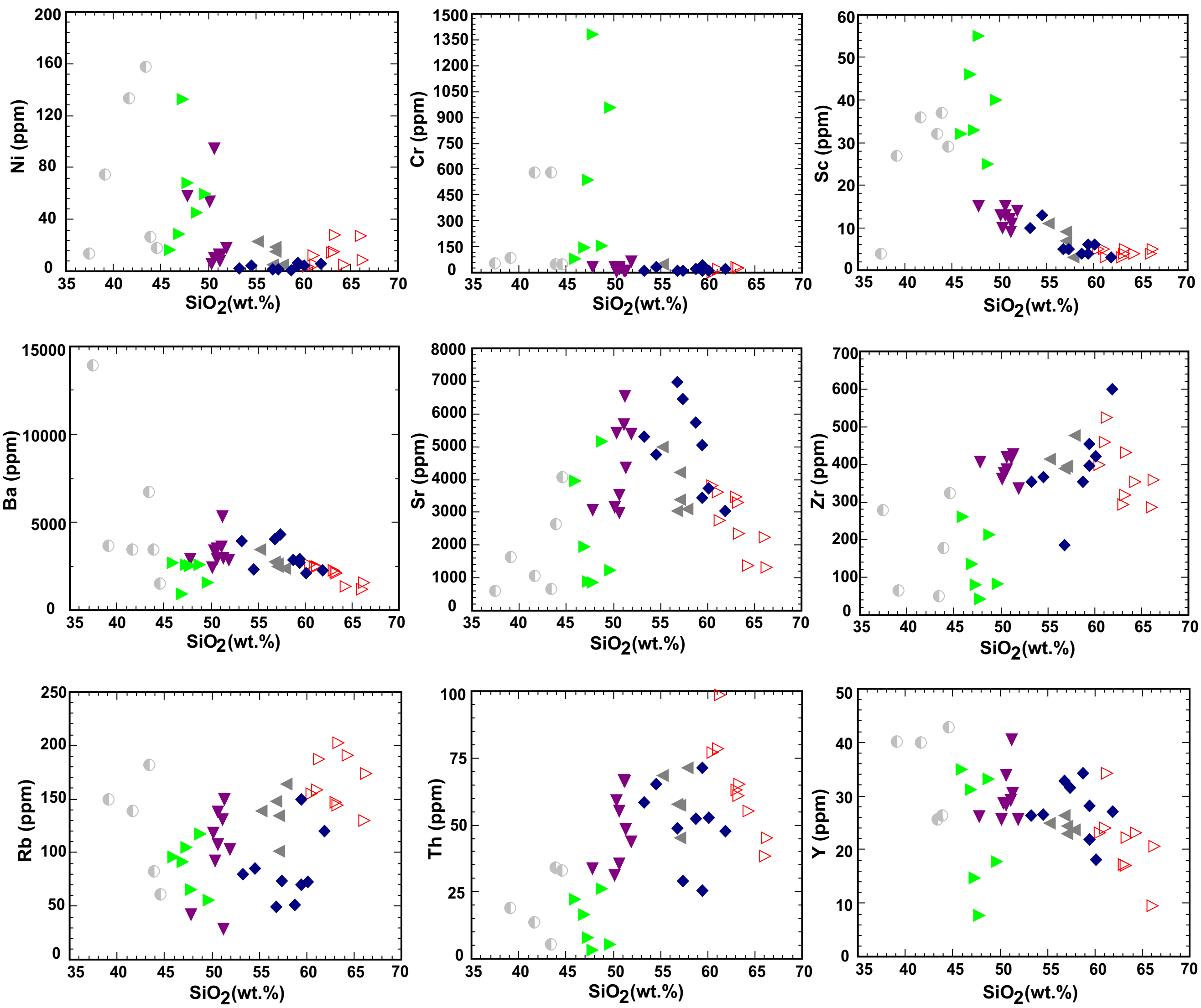

Fig. 8 Trace element Harker variation diagrams from Gölcük host rocks and xenoliths. Symbols are as in Fig. 4

and amphibole (hastingsite) minerals. Phlogopite and hastingsite are the main hydrous phases. Similarly, trachyte and trachyandesite, which are the host rocks, are found as phlogopite and hastingsite minerals as hydrous phase. The ultramafic, mafic, and felsic xenoliths, and their host rocks have crystallized from a relatively water-rich magma.

Textural features

Xenoliths are almost similar in terms of texture, and they are characterized by typically granular texture. Mineral phases do not show significant differences in the ultramafic, mafic, and felsic xenoliths and show a homogeneous distribution. For example, feldspar, clinopyroxene, and mica minerals exhibit almost the same grain size and texture features. Generally all minerals of xenoliths have not been influenced with corrosion, and they show anhedral to subhedral outlines. Textural features also support that the magma forming the rocks is not formed by the mixing (mingling) of magmas with different compositions. These textures emphasize that there is no change during the periods of mineral growth, but they emphasize that there is consistency with the progressive crystallization. The textural and mineralogical properties of ultramafic, mafic, and felsic xenoliths indicate that they were formed from magma forming the host rocks by fractional crystallization.

Geochemical interpretation

Xenoliths and host rocks of the Gölcük volcanism as illustrated by their variation diagrams, display SiO_2, CaO, and MgO contents varying from 37.5 to 66.3, from 1.41 to 18.21, and from 0.4 to 17.0 wt%, respectively. While the amounts of Na_2O,

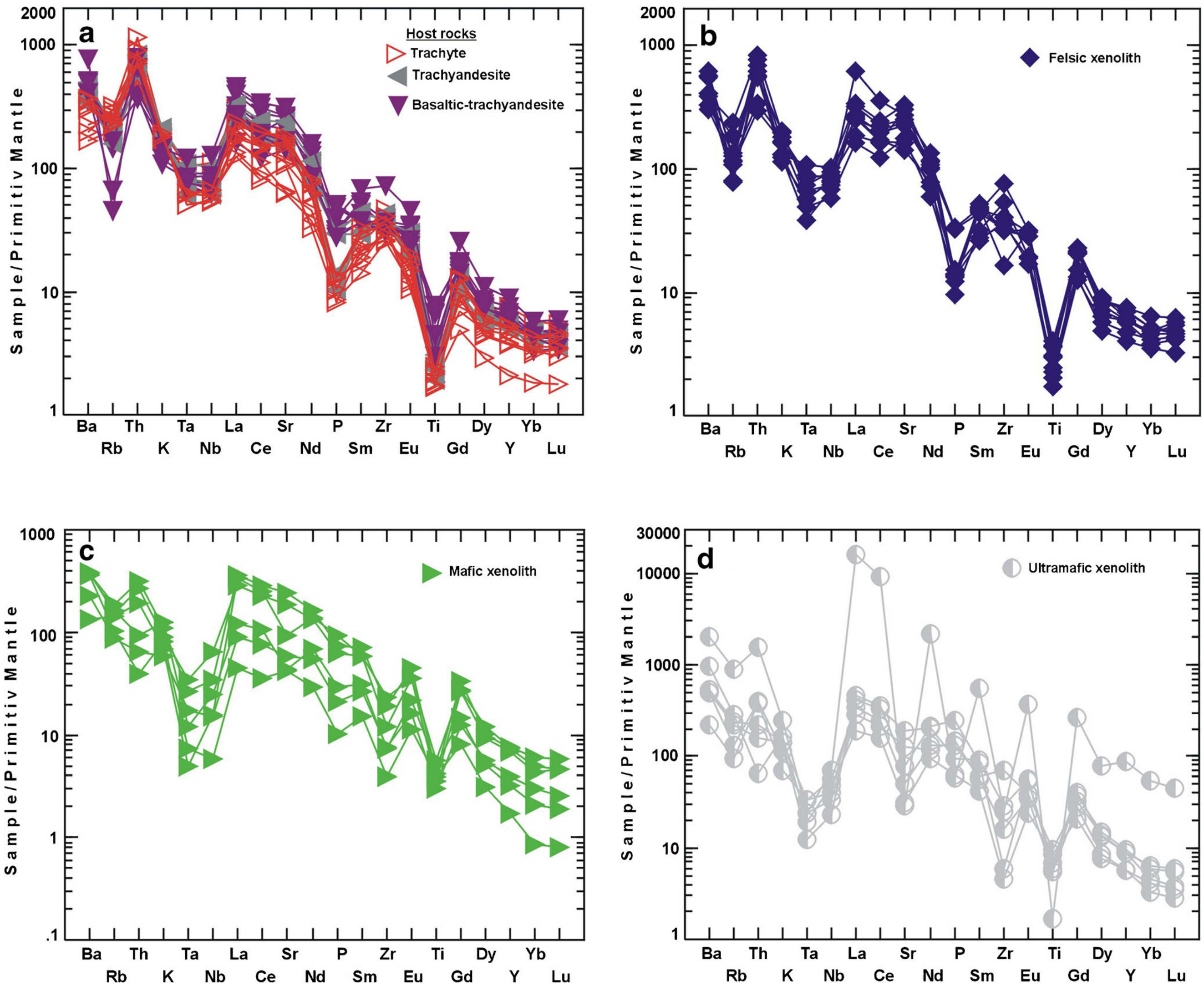

Fig. 9 Primitive mantle-normalized (normalization values from Sun and McDonough 1989) multielement variation diagrams of Gölcük host rocks and xenoliths. **a** Host rocks. **b** Felsic xenolith. **c** Mafic xenolith. **d** Ultramafic xenolith

K_2O, Sr, Th, Rb, and Zr increase with increasing SiO_2, it is seen that MgO, CaO, Fe_2O_3, TiO_2, MnO, P_2O_5, Ni, Sc, and Cr decrease. These variations play an important role in the crystallization of rocks of minerals such as plagioclase, clinopyroxene, amphibole, mica, apatite, and opaque minerals. MgO and CaO values are low in the host rocks and felsic xenoliths. High concentrations in the mafic and ultramafics indicate that there are not much ferromagnesian phases. They are more in the mafic and ultramafics xenoliths. Linear trends of MgO and CaO versus SiO_2 have been interpreted as the absence of mixing processes between xenoliths and parent rocks and possibly as a consequence of fractional crystallization. Ultramafic and mafic xenoliths from Gölcük have high Mg-numbers (63–91), high Ni (14–158 ppm), and high Cr (48–582 ppm), Mg-numbers (63–86), high Ni (17–133 ppm), and high Cr (82–1383 ppm), respectively, suggesting that they were derived from primitive upper mantle magmas. The low Mg-numbers (23–61), Ni (4–95 ppm), and Cr (4–68 ppm) contents of the host rocks and low Mg-numbers (35–67), Ni (0.9–6.4 ppm), and Cr (14–41 ppm) contents of the felsic xenoliths indicate a relatively evolved magma. Low Mg-numbers, low Cr, and Ni contents reflected by the abundant olivine and clinopyroxene phenocrysts were observed in samples that are related to crystallization in a magma.

All rocks are enriched in LILE. Primitive mantle-normalized spider patterns are characterized by enrichment of incompatible elements (Ba, Sr, Th, and K), while Nb, Ta, P, and Ti exhibit negative anomalies (Fig. 10). All studied rocks exhibit some Nb and Ta depletion typical of magmas generated from a subduction-metasomatized mantle (Wilson 2007). However, medium to high Zr contents against medium to low Nb contents of the Gölcük volcanics suggest strongly a subduction effect. Ba shows a strong and positive anomaly in incompatible elements of xenoliths. Enrichment in the main magma can be explained by aqueous fluid metasomatism in the mantle source containing phlogopite and/or amphibole. It

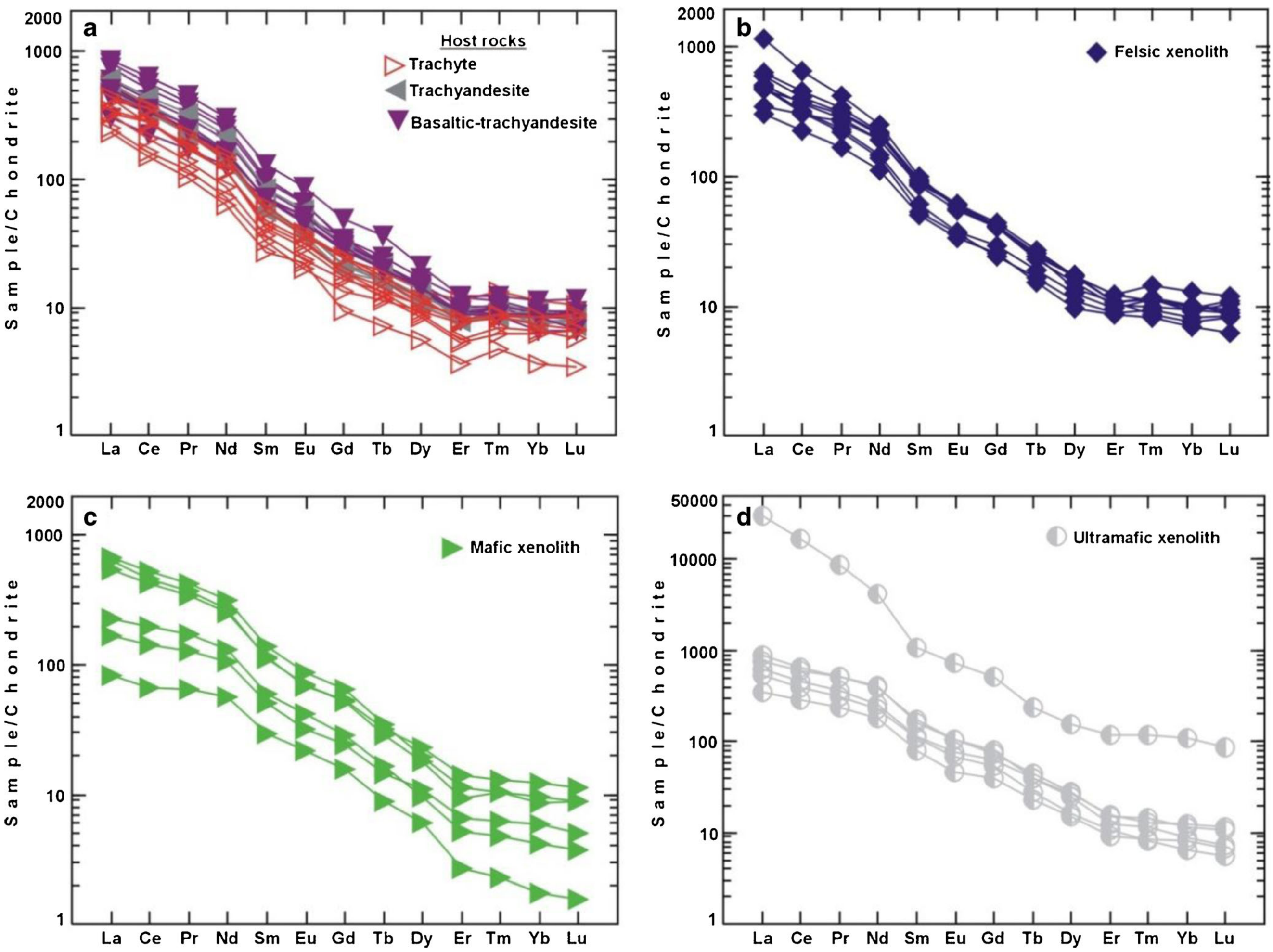

Fig. 10 Chondrite-normalized (normalization values from Taylor and McLennan 1985) REE variation diagrams from Gölcük host rocks and xenoliths. **a** Host rocks. **b** Felsic xenolith. **c** Mafic xenolith. **d** Ultramafic xenolith

is suggested that phlogopite and amphibole are the major reservoirs for LILE in the lithospheric mantle (Ionov et al. 1997).

A strong fractionation between LILE and HFSE combined with negative Nb, Ta, P, and Ti anomalies is commonly interpreted as indication of subduction processes involved in the petrogenesis of the magma (Thompson et al. 1984; Beccaluva et al. 1991). They may be attributed to the added of LILE released from the subducted oceanic crust (Pearce 1983; Hawkesworth et al. 1994). However, the enrichment of LILE and LREE relative to HREE and HFSE also reveals that a depleted mantle is metasomatized by fluids carrying LILE and LREE. All rocks display fractionated chondrite-normalized REE pattern with high and varying LREE concentrations in contrast to relative low HREE contents. High LREE/HREE ratios [$(La/Yb)_N = (29–97)$] reflect extremely high degrees of enrichment and show also steeper chondrite-normalized REE patterns. Strong enrichment in the LREE relative to HREE, presumably indicating the low-degree partial melts and presence of residual garnet in the source. Eu/Eu^* values for all the host rocks (except to Gl 29, Gl 39 samples) and xenoliths generally range from 0.82 to 1.20. The lack of a Eu anomaly suggests that plagioclase fractionation was not important during the origin and differentiation of the original magma.

In summary, petrographic features and various geochemical diagrams (Figs. 7, 8, 9, and 10) show that the Gölcük xenoliths compositionally resemble Gölcük host rocks, and they were formed from the same magmatic event. This situation is also interpreted that xenoliths and their host rocks were formed by fractional crystallization from a single magma.

Origin of the magma and xenolith interaction processes

This paper concerns the characteristics and origin of xenoliths and associated host rocks of the Gölcük volcanism. Petrographic, mineralogical, and geochemical similarities between xenolith and host rocks generally reflect a common source and same magma. Some hypotheses have been proposed to explain the origin of the alkaline magmatism in the study area. These include the differentiation of magmas generated from both metasomatised lithospheric and asthenospheric mantle sources (Coban 2003; Platevoet et al. 2014), a small degree of partial melting of metasomatised lithospheric mantle (Alıcı et al. 1998; Elitok et al. 2010), and fractional crystallization (Alıcı et al. 1998; Hildenbrand et al. 1999) and (negligible) assimilation of continental crust (Alıcı et al. 1998).

Yılmaz (2010) emphasized that lithospheric and sub-lithospheric mantle melts play an important role in the origin of Denizli potassic lava, which is located in the same parallel with Gölcük volcanism. It is noteworthy that mantle-derived volcanic rocks showing OIB-like geochemical features are commonly found in this orogenic setting (Aldanmaz et al. 2006; Dilek and Altunkaynak 2007, 2008). Caran (2016) reported that carbonatitic and kamafugitic melts formed by the partial melting of carbonatite mantle metasomatism have effects at the origin of silica-undersaturated potassic magmas near the Gölcük volcanism.

The similar mineralogical and geochemical characteristics of the host rocks and xenolites in Gölcük provide evidence for stages of partial assimilation. The ultramafic xenoliths are composed mainly of clinopyroxene (diopside) and mica (phlogopite), while mafic ones are formed from basic plagioclase together with these dark-colored minerals. These xenoliths support the existence of deeper magma chamber where clinopyroxene, mica (phlogopite), and basic plagioclase were fractionating phases. Felsic xenoliths have been incorporated in the magma chamber at shallow depth during magma rise and are composed of alkali feldspar, sodic plagioclase, phlogopite, diopside, and amphibole minerals. The dark-colored xenoliths (basic and ultrabasic) are from minerals with high melting temperature. Felsic xenoliths are composed of minerals with a lower melting temperature. They have been incorporated in the magmas at shallow depths during magma rise. Therefore, felsic xenoliths are being subject to magma influence for a short time than basic and ultrabasic xenoliths, and according to this, the assimilation of dark-colored xenoliths was occurred at a longer time. Although they have been in the magma for a long time, assimilation is not much more advanced. Although the felsic xenoliths were found to have short-term contact with magma, the effect of assimilation might have been advanced due to the fact that the minerals forming (e.g., alkali feldspar, sodic plagioclase) were more melting. In addition, the magma did not have a high temperature for a long enough period for assimilation of felsic xenoliths and crustal rocks. Because the host rocks contain sparsely claystones that have not undergone alteration under the influence of the parent magma. It is interpreted that the magma was not at such high temperature to melt the claystones or these rocks were not in contact with magma for a sufficiently long time.

The study of xenoliths and their host rocks provided evidence for fractional crystallization as the main rock-forming processes in the Gölcük. Fractional crystallization is shown by consistent trends of some components on the Harker diagrams. In general, all xenolith types exhibit similar trends as the host rocks. Negative trends of Fe_2O_3, MgO, CaO, and TiO_2 with silica can be expressed as fractionation of Ca, Mg, and Fe-rich minerals during crystallization of magma. Clinopyroxene, phlogopite, and amphibole minerals are the most suitable for such crystallization as they contain the mentioned elements. Clinopyroxene and phlogopite are most concentrated in ultramafic and mafic xenoliths, whereas amphibole occurs mostly in felsic xenoliths and host rocks. Al_2O_3 flat correlation with silica in felsic xenoliths and host rocks may indicate a minor fractionation phase of plagioclases. Plagioclase fractionation can be said to be more important with increasing the acidity of the residual melt. The distribution of SiO_2 versus Na_2O can be attributed to sodic plagioclases fractionation from low silica melts. Amphibole is the main mineral to fractionate, for host rocks and felsic xenoliths, as shown by the slope of CaO, MgO, and Fe_2O_3. The oxidation of the rims of the amphiboles found in these rocks indicates that they are crystallized at low temperature and pressure at during magma ascent.

The negative tendency of MgO can be interpreted as the main mineral phase, of phlogopite and clinopyroxene at dark-colored xenoliths. Crystallization of mafic minerals (diopside, phlogopite) at moderate to high pressures than plagioclase may be considered as a main process for the evolutionary trends of magma (Grove and Baker 1984). Olivine does not find as phenocrysts in other rocks except for the lamprophyric dykes in the Gölcük volcanism (Platevoet et al. 2014). This fact as well as the scattered distribution of Cr and Ni make olivine fractionation during differentiation of basic and ultrabasic xenolithes to be unlikely. Diopside and phlogopite-bearing mafic, ultramafic xenoliths may be interpreted as the formation of fractional crystallization in the deeper magma chamber.

Conclusions

Alkaline, ultrapotasic, and shoshonitic magmatism occur in various geological setting, generally in volcanic arc setting above subduction zones. The dominant main rocks (host rocks) are trachyte, trachyandesite, basaltic trachyandesite and with small volumes of phonolite, tephrite and lamprophyres, and pyroclastics. They occur as dykes and domes with country rocks. Within the Gölcük, three types of xenoliths with ultramafic (pyroxenitic), mafic (gabbroic, monzo-gabbroic), and felsic (syenitic, siyeno-dioritic) composition were identified. They are occasionally elliptical and rounded to angular forms, several tens of centimeters in diameter, fine- to medium-grained, and holocrystalline. Mineral chemistry of main rocks and xenoliths in the Gölcük is studied in detail by Platevoet et al. (2014) and Elitok et al.

(2010). Clinopyroxene (diopside, augite) and Mg-rich mica (phlogopite) were the main fractionating phases in ultramafic xenoliths. The mafic xenoliths were fractionated with clinopyroxene (diopside, augite), mica (phlogopite), plagioclase (between An_{42} and An_{19}), and apatite. The felsic xenoliths composed mainly of K-feldspar (anorthoclase or sanidine), plagioclase (between An_{42} and An_{19}), clinopyroxene (diopside, augite), mica (phlogopite), amphibole (magnesio-hastingsite), and apatite.

The felsic xenoliths have chemical signatures almost similar to their host rocks (trachyte, trachyandesite). They are depleted in MgO, CaO, Fe_2O_3, TiO_2, P_2O_5, Cr, Ni, and Sc and enriched in SiO_2, K_2O, and Na_2O. Ultramafic and mafic xenoliths have high concentrations of Fe_2O_3, CaO, MgO, Y, and transition elements (Cr, Ni); and low concentrations of K_2O, Na_2O, P_2O_5, and Zr. Geochemically, all rocks of the Gölcük volcanics (host rocks and xenoliths) were enriched in LILE (e.g., K, Th, Ba, Sr) and depleted in HFSE (e.g., Nb, Ta, Ti, P). The chondrite-normalized REE patterns are characterized by LREE enrichment [$(La/Yb)_N = (28.5–97.4)$]. Geochemical results of this study combined with previously published data on the Gölcük indicate that magmas produced by partial melting of mantle sources modified by metasomatic processes associated with a subducted slab. Metasomatized lithospheric mantle was homogeneously fedded with small volume asthenospheric melts. Mantle-source melts have taken place in deep-seated magma chambers then undergone a processes of evolution involving fractional crystallization during their ascent to the surface. In other words, host rocks and different types of xenoliths (ultramafic, mafic and felsic) were controlled by differentiation processes.

The main findings for the main rocks and xenoliths in Gölcük can be summarized as follows:

1. Xenoliths hosted by pyroclastic deposits and lavas from Gölcük are undeformed fragments of ultramafic (pyroxenitic), mafic (gabbroic and monzo-gabbroic), and felsic (syenitic and siyeno-dioritic) plutonic rocks.
2. Xenoliths are seen as individual fragments in the lavas and piroclastics deposits. They are found more in pyroclastics deposits rather than lavas.
3. Mineral assemblages of xenoliths are generally similar to the host rocks, differing only in their proportions. The similarity of mineralogical and geochemical (major and trace element) compositions of the all xenoliths types and host rocks suggests that these have a similar origin.
4. Xenoliths formed by fractional crystallization from the single magma emplaced at crustal depth as their host rocks.

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