Polygenetic olivine phenocrysts in Okete basanite, New Zealand

NINOMIYA Atusi and Shoji ARAI

*Department of Earth Sciences, Kanazawa University,
Kanazawa 920-1192, Japan*

Generation of polygenetic olivine phenocrysts is discussed for a basanite from Okete Quarry, New Zealand, based on descriptions of its olivine and spinel and the reaction product between the basanite and its peridotite xenoliths. The olivine phenocrysts can be divided into two types, I and II. The phenocrysts (I) are relatively small, subhedral to euhedral and has minute spinel inclusions sparsely throughout the grain. They are free of the textures indicating deformation and fracture. The phenocrysts (II) are relatively large and have spinel inclusions only in the rim. Their core has deformation textures and/or fluid inclusion trails. The olivine phenocrysts (I) are normally zoned: the core ranges from Fo_{85} to Fe_{76} and the rim has Fe_{76} . The spinel inclusions are highly variable in $Cr/(Cr+Al+Fe^{3+})$ atomic ratio (0.02–0.5), $Cr\# (=Cr/[Cr+Al])$ atomic ratio (0.1–0.6), TiO_2 content (17–1 wt%) and $Fe^{3+}/(Cr+Al+Fe^{3+})$ atomic ratio ($=Fe^{3+}$ ratio) (0.7–0.2), and can be grouped into two, low- Fe^{3+} (Fe^{3+} ratio <0.5) and high- Fe^{3+} (Fe^{3+} ratio >0.5) spinels. The Cr ratio is correlated positively with the $Cr\#$ and negatively with the TiO_2 content and Fe^{3+} ratio. The phenocrysts (II) frequently have compositionally flat cores, which are relatively Mg-rich (up to Fe_{90}), and zoned rims (Fe_{85-76}). Their spinel inclusions mostly belong to the high- Fe^{3+} group. The generation of olivine crystals and their spinel inclusions of various origins can be observed at the boundaries between the basanite and its peridotite xenoliths. Orthopyroxene in the peridotite xenoliths was decomposed into olivine, clinopyroxene, spinel and melt (glass) by the reaction with the basanite melt. The spinel from the reaction product is relatively high- $Cr\#$. Subhedral olivine which includes chromian spinels with a wide compositional range has overgrown on the mantle olivine, which has been gradually disintegrated into the basanite matrix.

The olivine phenocrysts (I) with low- Fe^{3+} spinel inclusions were derived from the orthopyroxene/melt reaction product, and those with high- Fe^{3+} spinels were precipitated from the basanite melt. The olivine phenocrysts (II) are of xenocrystal origin and have been modified by the melt to various degrees. The olivine phenocrysts in olivine-rich magmas such as picrite basalts and alkali olivine basalts can be polygenetic due to the reaction between the mantle peridotite and more primitive magmas.

Keywords: Basanite, Peridotite xenolith, Okete, New Zealand, Polygenetic olivine phenocrysts, Spinel, Peridotite/melt reaction

I. Introduction

What are the phenocrysts in volcanic rocks? It has been known that the phenocrysts are quite complicated in their origin. For example the magma mixing could produce complicated phenocryst assemblages and chemistries (e.g., Sakuyama, 1978, 1984). Considering the complicated processes that magmas

possibly have experienced in deep parts, such as magma mixing, magma/wall rock interaction, fractional crystallization and assimilation, the phenocrysts are possibly polygenetic. The polygenetic nature of the olivine phenocrysts is examined for a basanite from Okete Quarry, New Zealand, in this article. The term “phenocrysts” is used to indicate the “large megascopically visible crystals” according to

Fig. 1. Locality map of Okete Quarry, New Zealand (after Briggs and Goles, 1984).

the classical definition of Iddings (e.g., Holmes, 1928). Boudier (1991) and Ninomiya and Arai (1993) interpreted that olivine phenocrysts in picrite are polygenetic as the product of reaction between basaltic melt and mantle harzburgite (e.g., Kelemen, 1990; Arai *et al.*, 1994) at relatively low pressures (the uppermost mantle to lower crust). One of the advantages for the study of Okete basanite is that the process of formation of polygenetic olivine phenocrysts can be directly observed at boundaries between the basanite magma and its peridotite xenoliths. Preliminary observations on the contribution of the reaction between magmas and orthopyroxene of mantle peridotite xenoliths, including the Okete peridotite-basanite pair, were reported by Arai *et al.* (1994). In this study, 19 thin sections for basanite-xenolith pairs from Okete Quarry, New Zealand, were examined under the microscope and 16 of them were analyzed for mineral chemistry.

II Geological background

The basanite of the Okete Quarry, western North Island, New Zealand, belongs to the Okete Volcanics (or Okete Volcanic Formation) of the Alexandra Volcanics of Plio-Pleistocene (Briggs, 1983; Briggs and Goles, 1984) (Fig. 1). The Okete Volcanics are composed of many small monogenetic volcanoes which are made mostly of pyroclastics with a few thin lava flows. They are alkalic in bulk composition, ranging from basanite, alkali basalts to hawaiite while the rest of the Alexandra Volcanics are subalkalic (Briggs, 1983). The samples examined in this study were taken from the Okete Quarry, the type locality of the Okete Volcanics (R14/788750; grid of 1:50,000 topographical map series of New Zealand). The age of the Okete Quarry basanite was reported to be 1.80 or 2.69 Ma (Robertson, 1976). The Okete basanite and alkali basalt contain abundant peridotite xenoliths (e.g., Rodgers *et al.*,

1975). They are variable in lithology, from peridotites (dunite, harzburgite, lherzolite and wehrlite) to clinopyroxenites (Rodgers *et al.*, 1975). The Okete basanites contain 44 wt% of SiO_2 , 2.2 wt% of TiO_2 , and 12.0 wt% of MgO on average (Briggs and Goles, 1984). Their K_2O and Na_2O contents and ratio of $\text{K}_2\text{O}/\text{Na}_2\text{O}$ are 1.2 wt% and 2.7 wt% and 0.44 on average, respectively (Briggs and Goles, 1984). The FeO (as total Fe)/ MgO weight ratio varies from 0.49 to 0.60. The Okete basanites contain 8.7 wt% of normative nepheline on average (Briggs and Goles, 1984). Briggs and Goles (1984) suggested that the Okete basanite magma was derived from a light REE-enriched garnet peridotite at above 2 GPa.

III. Petrographical descriptions

1. Basanite

The basanite (alkali basalt) from the Okete Quarry has an intersertal texture and is highly olivine-phyric. Phenocrysts of olivine occupy 16 vol. %. Phenocrysts of clinopyroxene (ca. 1 vol. %) and rare plagioclase are also found. The groundmass is composed of clinopyroxene, plagioclase, Ti-magnetite and clear brown glass. Modal nepheline is absent.

The olivine phenocrysts are quite variable in morphology and other petrographical characteristics. In this paper the olivine phenocrysts are classified into two types, I and II. The olivine phenocrysts (I) have euhedral to subhedral forms and are free of textures indicating deformation (kink band and undulatory extinction) and fracture (fluid inclusion trails) (Fig. 2a). They often include minute euhedral spinels sparsely throughout the grain. The olivine phenocrysts (II), on the other hand, have the deformation and fracture textures in the core (Figs. 2bc and 3). The minute spinels are included only in the rim, which is sometimes distinctively Fe-rich (Fig. 3). The size is slightly different between the two types: the

Fig. 2. Photomicrographs of olivine phenocrysts in the Okete basanite. (a) Highly euhedral olivine phenocryst (I). Note that the spinel inclusions (arrows) are even in the inner part of the phenocryst. Plane-polarized light. Scale bar is 0.1 mm. (b) Subhedral olivine phenocryst (II) preserving a kink band. Note that spinel inclusions (black dots) (arrows) are only in the marginal part of the phenocryst. Crossed-polarized light. Scale bar is 0.5 mm. (c) Anhedral olivine phenocryst (II) showing kink bands. Note the morphological modification (embayment and subhedral overgrowth) (see Fig. 3). Crossed-polarized light. Scale bar is 0.5 mm.

Fig. 3. SEM micrographs of the olivine phenocryst (II). (a) Back scattered electron image of the olivine phenocryst (II) of Fig. 2c. The olivine has a compositionally flat Mg-rich (dark) core and a light gray Fe-rich rim. (b) A close-up of a marginal part of (a) (rectangle). Note that spinel inclusions (bright spots, arrow) are only in the rim which is partly euhedral.

averaged size is 0.12 mm long and 0.19 mm long for phenocrysts (I) and (II), respectively. Type I (82%) surpasses type II (18%) in volume in the total olivine phenocrysts. The color of spinel inclusions is variable in thin section from light brown to black (opaque), indicating a wide chemical variation.

2. Peridotite xenolith

The Okete basanite contains small peridotite xenoliths which are usually a few centimeters across and range from clinopyroxene-

poor lherzolite to clinopyroxene-bearing harzburgite with weakly porphyroclastic textures. Olivine grains are usually anhedral and deformed. Chromian spinel grains are irregular in shape and are green to brown through light brown in thin section, indicating low to intermediate $Cr\#s (=Cr/(Cr + Al)$ atomic ratio). Orthopyroxene porphyroclasts have clinopyroxene lamellae.

3. Peridotite xenolith rim

All peridotite xenoliths examined here have considerably complicated textures in their rim, indicating an intense reaction with the host basanite magma which accompanies mechanical disintegration of peridotite olivine.

Olivine grains in direct contact with the host basanite show relatively fine grained subhedral morphology (Fig. 4a). The abundance and thickness of the subhedral part of olivine are variable along the xenolith/basanite boundary (Fig. 4cd). The subhedral part has euhedral spinel inclusions indicating its overgrowth origin from the host basanite (Fig. 4a). The size of the subhedral olivine on the mantle olivine in the xenolith rim sometimes seems to be constrained by the width of the subgrain of olivine in the xenoliths (Fig. 4a). Several olivine grains which are fine and rounded irregular in shape frequently make a cluster in the xenolith rim, especially in the embayed part. This type of modification sometimes proceeded deeply into the xenolith interior by invasion of the melt, resulting in splitting the xenolith into smaller pieces (Fig. 4b). In this case, relatively large olivine grains sometimes with deformation textures (=mantle peridotitic pieces) are surrounded by spinel-bearing subhedral olivine grains (Fig. 4b).

The orthopyroxene in the peridotite xenoliths in direct contact with the host basanite is always intensely decomposed into olivine aggregates with interstitial clinopyroxene, glass, plagioclase and rare apatite (Fig. 5a-

Fig. 4. Photomicrographs of subhedral olivines around the rim of peridotite xenoliths. (a) Subhedral olivine with spinel inclusions (arrow) overgrowing on xenolithic olivine without spinel inclusions. Note that grain size of the subhedral part is similar to the width of the subgrain. Crossed-polarized light. Scale bar is 0.5 mm. (b) A peridotite fragment surrounded by subhedral olivine grains with minute spinel inclusions (black dots). It seems to be disintegrated from the xenolith (lower) due to infiltration of basanite melt. Plane-polarized light. Scale bar is 1 mm. (c) Peridotite xenolith rim (lower) with thick overgrowth of subhedral olivine. Plane-polarized light. Scale bar is 1 mm. (d) Peridotite xenolith rim with thin overgrowth of subhedral olivine. Note that (c) and (d) are from the same sample. Plane-polarized light. Scale bar is 1 mm.

d). This is ascribed to the reaction between orthopyroxene and basanite groundmass because orthopyroxene is decomposed only when it is in contact with the basanite magma. The interstitial material varies outward from clinopyroxene (orthopyroxene side) (Fig. 5b) through glass (Fig. 5c) to basanitic matrix (Fig. 5d). The olivine is relatively fine-grained, much finer than the olivine grains in the peridotite, and is anhedral (Fig. 5b) to rounded subhedral with fine euhedral spinel inclusions (Fig. 5cd). Neighboring several fine-grained olivines often share the same crystallographic

orientation, possibly being a part of a highly anhedral or skeletal single olivine crystal (e.g., Fig. 5b). Orthopyroxene may have sometimes reacted out and the olivine grains appear to be released from xenolith into the host magma (Fig. 5e). In such a case olivine is a little more coarsened than in earlier stages of reaction. The fine olivine grains, the reaction product, are approximately comparable in size (0.12 mm long on average) and shape with the olivine phenocrysts (I) in the basanite (Figs. 2a and 5). The decomposition of orthopyroxene may also promote the disintegration of the peridotite

Fig. 5. Photomicrographs of olivines in the orthopyroxene/basanite reaction product. (a) A reaction zone between basanite (right) and orthopyroxene (opx). Circles are scopes of (b), (c), and (d) below. Crossed-polarized light. Scale bar is 1 mm. (b) Inner part of the reaction zone adjacent to orthopyroxene comprising anhedra olivine (ol) and clinopyroxene (cpx). The clinopyroxene with cleavage poikilitically includes subhedral olivine. Crossed-polarized light. Scale bar is 0.1 mm. (c) Central part of the reaction zone. Subhedral olivine (ol) and spinel (sp) grains with colorless interstitial glass (gl). Plane-polarized light. Scale bar is 0.1 mm. (d) Outer part of the reaction zone adjacent to the basanite. Basanitic matrix (mat) is interstitial to subhedral olivine with spinel. The basanite host is towards the upper left (arrow). Plane-polarized light. Scale bar is 0.1 mm. (e) Cluster of subhedral olivine grains of possible reaction origin in loose contact with the rim of xenolith (lower right). Note that some of the olivine grains (arrow) seem to be released from the cluster. Crossed-polarized light. Scale bar is 0.5 mm. (f) The alteration of chromian spinel and disintegration of olivine in the peridotite xenolith rim accompanying with the orthopyroxene/basanite reaction. Plane-polarized light. Scale bar is 1 mm.

xenolith, assisting the release of the peridotite olivine into the host basanite (Fig. 5f).

Clinopyroxene in the xenolith rim usually shows a spongy appearance due to decompressional partial melting (e.g., Donaldson, 1978) and is slightly overgrown by clear clinopyroxene in contact with the host basanite. Chromian spinel in the xenolith rim is decomposed into aggregates of minute euhedral dark spinel phase, preserving the outline of the primary one (Fig. 5f).

IV. Mineral chemistry

Minerals were analyzed by a scanning electron microscope, Akashi alpha-30A, with an energy-dispersive spectrometer at Kanazawa University. Analyzing condition was 20 kV accelerating voltage, 30° tilting angle, 30 mm working distance and 50 mm detector-to-centerline distance for SEM, and 18 kV accelerating voltage, 30° tilting angle, 38.5° take-off angle and 1.0 beam current constant for EDAX. 100-second integration time was adopted for each analysis. Natural and synthetic minerals were used as standards. All Fe in silicates is assumed to be ferrous. Fe^{2+} and Fe^{3+} in spinels were calculated assuming spinel stoichiometry after all Ti was subtracted as ulvöspinel molecule (Fe_2TiO_4). The Mg# in spinel phase indicates the $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ atomic ratio for $(\text{Mg}, \text{Fe}^{2+})(\text{Cr}, \text{Al}, \text{Fe}^{3+})_2\text{O}_4$ molecules in this paper. Selected analyses are listed in Table 1.

1. Peridotite xenoliths

The peridotite xenoliths examined in this study are plotted within the olivine-spinel mantle array, a spinel peridotite restite trend in terms of the Fo content of olivine and the Cr# ($=\text{Cr}/(\text{Cr} + \text{Al})$ atomic ratio) of chromian spinel (Arai, 1987, 1994a). The Fo content of olivine is around Fo_{90-91} (Fig. 6) and the Cr# of spinel is within a range from 0.1 to 0.4, concordant with the petrographical observations mentioned

above (Fig. 6) (Arai, 1994a). Fe^{3+} and Ti contents of primary spinels are very low (Table 1; Figs. 7 to 9).

2. Olivine and its spinel inclusion in basanite

The olivine phenocrysts (I) show the normal compositional zoning: the rim composition is around Fo_{77} and the core composition varies from Fo_{85} to Fo_{77} . On the other hand, their spinel inclusions, which are chemically homogeneous, are highly variable in composition depending on the host olivine, that is, show an approximate inter-grain homogeneity in individual olivine grains. The $\text{Fe}^{3+}/(\text{Cr} + \text{Al} + \text{Fe}^{3+})$ atomic ratio ($=\text{Fe}^{3+}$ ratio) of the spinel inclusions varies from 0.2 to 0.7 (Fig. 7), positively proportional to the TiO_2 content, which varies from 1 to 17 wt% (Fig. 9). The Cr# is positively correlated with Ti and Fe^{3+} contents (Figs. 8 and 9). The spinel inclusions are apparently divided into two groups; low- Fe^{3+} (Fe^{3+} ratio < 0.50) and high- Fe^{3+} (Fe^{3+} ratio > 0.50) (Fig. 7). The Mg# of spinels is lower in the low- Fe^{3+} group than in the high- Fe^{3+} group (Fig. 10).

The olivine phenocrysts (II) generally show normal chemical zonings, being Fo_{77} at the very rim. The core composition is possibly dependent on the grain size: large grains (ca. > 1 mm across) have compositionally flat Mg-rich core ($\sim \text{Fo}_{90}$) and chemically zoned rim, around 50μ in width (Fig. 3). The spinel inclusions, which are only in the rim of the olivine phenocrysts (II), are almost high- Fe^{3+} ratio (0.25–0.72) and low-Cr# (0.12–0.54), the Fo content of coexisting olivine varying from 77 to 85 (Fig. 6). The spinel inclusions in the phenocryst (II) rims are similar in chemistry to the high- Fe^{3+} group spinel included in the phenocrysts (I) (Figs. 7 to 10).

3. Olivine and spinel in xenolith rim

TiO_2 content and Fe^{3+} ratio of spinel grains associated with olivine of the orthopyroxene/melt reaction product, which are nega-

Fig. 6. Relationships between Cr/(Cr+Al) atomic ratio of spinel and Fo content of coexisting olivine from the Okete basanite and its peridotite xenoliths. Olivine-spinel pair of each xenolith is an average for 2 to 4 analyses. Note that the Fo content of olivine is not correlated with the Cr# of coexisting spinel. OSMA, olivine-spinel mantle array, which shows the compositional field of spinel peridotite restite (Arai, 1987, 1994a). Phen. I & phen. II, olivine phenocrysts (I) & (II), respectively. Low-Fe³⁺, high-Cr# and high-Fe³⁺, low-Cr#; chemical classification of spinel inclusions in olivine. Arrow indicates an approximate variation outward in the orthopyroxene/basanite reaction product (see text).

tively correlated with the Cr#, increase outward toward the matrix. The spinel is relatively high in Cr#, 0.40–0.72, showing a weak positive correlation with the Fo content of host olivine (80.0–89.1) (Fig. 6). Both the Fo of olivine and the Cr# of spinel in the orthopyroxene/melt reaction product decrease outward to the basanite matrix (Fig. 6). This Fo–Cr# trend is continuous to the olivine–low-Fe³⁺ spinel trend for the phenocrysts (I) (Fig. 6). The Mg# of spinel formed by the orthopyroxene/melt reaction ranges from 0.47 to 0.70 (Fig. 10).

The Fe-rich rim including the subhedral overgrowing part of the olivine in the xenolith rims in contact with basanite is wider (ca. 0.6 mm) than the rim of the olivine phenocrysts (II) (Fig. 4). Spinel grains included in the subhedral part show a wide range in chemistry: the Fe³⁺ ratio (0.15–0.70) and the Cr# (0.15–0.60) show negative and positive relationships, respectively, with the TiO₂ content (2.7–16.5 wt%) (Figs. 8 and 9). The Mg# of spinel ranges from 0.46 to 0.65 (Fig. 10). It is noteworthy that the compositional range of spinel in the subhedral olivine on the xenolith rim roughly covers the whole compositional ranges of

spinel in the phenocrysts (I) and the rim of the phenocrysts (II) (Figs. 7 to 9).

V. Discussion: formation of polygenetic olivine phenocrysts in the basanite

The wide textural and compositional variations of olivine as well as spinel from the Okete basanite strongly indicate that olivine phenocrysts in magmas can be polygenetic due to the reaction or interaction between mantle peridotite and magma, including the related disintegration of peridotite. The reaction or interaction process observed in the Okete basanite–peridotite xenolith system is analogous to the interaction between wall peridotite and magma at Moho or uppermost mantle condition (e.g., Kelemen *et al.*, 1990) because the reaction between Mg-rich orthopyroxene and magma with precipitation of olivine is common to both the processes.

1. Three kinds of euhedral olivine crystals

Three kinds of olivines can be formed during the reaction between peridotite and

Fig. 7. Ratios of trivalent cations of spinels in the basanite and its peridotite xenoliths from Okete, New Zealand. Note the linear compositional trend for almost all spinels in the basanite. Abbreviations are as described in Fig. 6. Arrow indicates an approximate variation outward in the orthopyroxene/basanite reaction product (see text).

Fig. 8. Relationships between TiO₂ wt % and Cr/(Cr+Al) atomic ratio of spinels from the Okete basanite and its peridotite xenoliths. Note the linear compositional trend for almost all spinels of the basanite. Abbreviations are as described in Fig. 6. Arrow indicates an approximate variation outward in the orthopyroxene/basanite reaction product (see text).

magma and associated disintegration of the peridotite. The first is a reaction of orthopyroxene in the peridotite: orthopyroxene + melt(1) = olivine + clinopyroxene + melt(2), is widely observed under the microscope (Fig. 5). This olivine is rounded subhedral and relatively fine-grained, occasionally having minute spinel

inclusions (Fig. 5c-e). The second is a xenocrystal olivine which has been released from the peridotite xenoliths and has been modified (overgrowth and dissolution) by the host melt to be euhedral to various extent (Fig. 2bc). The degree of the morphological modification is various, from being almost intact (totally an-

Table 1. Selected microprobe analyses of olivine and spinel from the Okete basanite and its peridotite xenolith

mineral	peridotite xenolith								basanite			
	olivine	spinel	opx	cpx	overgrowth rim olivine	spinel	opx-melt reaction olivine	spinel	phenocryst (I) olivine	spinel	phenocryst (II) olivine	spinel
No.	OQ4306	OQ4304	OQ7101	21314cp	OQ7403	OQ7404	1145Solreac	1146spreac	OQ4410	OQ4411	ol102c	sp102b
SiO ₂	41.27	0.23	57.19	53.20	39.59	0.36	39.89	0.01	39.32	0.65	38.88	0.59
TiO ₂	0.00	0.39	0.00	0.31	0.05	14.26	0.00	9.19	0.15	1.78	0.10	15.60
Al ₂ O ₃	0.28	38.10	2.31	3.15	0.27	8.26	0.00	15.24	0.29	17.10	0.16	7.24
Cr ₂ O ₃	0.04	25.45	0.42	1.04	0.03	4.65	0.06	19.24	0.16	32.34	0.00	2.39
FeO*	9.56	15.62	5.74	3.20	17.85	59.32	15.28	39.61	17.52	34.11	20.42	62.69
MnO	0.14	0.00	0.12	0.21	0.31	0.45	0.34	0.02	0.44	0.10	0.29	0.38
MgO	48.71	17.19	33.76	15.80	41.43	7.32	43.25	9.45	42.34	9.24	38.97	5.97
CaO	0.08	0.17	0.51	22.60	0.45	0.11	0.37	0.21	0.35	0.19	0.49	0.22
Na ₂ O	0.00	0.00	0.00	0.16	0.00	0.00	0.00	0.72	0.00	0.00	0.00	0.00
K ₂ O	0.03	0.04	0.00	0.00	0.00	0.05	0.00	0.07	0.02	0.01	0.01	0.06
NiO	0.37	0.25	0.05	0.00	0.31	0.14	0.39	0.37	0.17	0.22	0.17	0.23
total	100.48	97.44	100.10	99.67	100.29	94.92	99.58	94.13	100.76	95.74	99.49	95.37
Number of atoms												
O	4.	4.	3.	6.	4.	4.	4.	4.	4.	4.	4.	4.
Si	1.005	0.006	0.981	1.940	1.005	0.013	1.009	0.003	0.994	0.022	1.005	0.023
Ti	0.000	0.008	0.000	0.008	0.001	0.414	0.000	0.245	0.002	0.046	0.002	0.456
Al	0.008	1.302	0.046	0.135	0.008	0.376	0.000	0.638	0.008	0.698	0.004	0.332
Cr	0.000	0.583	0.005	0.030	0.000	0.142	0.001	0.540	0.003	0.886	0.000	0.073
Fe*	0.194	0.378	0.082	0.097	0.379	1.917	0.323	1.177	0.370	0.988	0.441	2.040
Mn	0.003	0.000	0.006	0.006	0.009	0.020	0.007	0.000	0.009	0.003	0.006	0.012
Mg	1.770	0.747	0.863	0.859	1.568	0.421	1.630	0.500	1.594	0.477	1.501	0.346
Ca	0.002	0.005	0.009	0.883	0.012	0.004	0.010	0.008	0.009	0.007	0.013	0.009
Na	0.000	0.000	0.000	0.011	0.000	0.000	0.000	0.049	0.000	0.000	0.000	0.000
K	0.001	0.001	0.000	0.000	0.000	0.002	0.000	0.003	0.000	0.000	0.000	0.003
Ni	0.007	0.005	0.000	0.000	0.006	0.004	0.008	0.010	0.003	0.006	0.003	0.007
total	2.990	3.035	1.992	3.969	2.988	3.313	2.988	3.173	2.992	3.133	2.975	3.301
Mg#	0.901	0.745	0.913	0.899	0.805	0.623	0.835	0.634	0.812	0.326	0.773	0.552
Cr		0.293				0.105		0.342		0.449		0.058
Al		0.653				0.278		0.405		0.354		0.265
Fe		0.054				0.617		0.253		0.196		0.677
Cr#		0.309				0.274		0.458		0.559		0.180
Fo		90.1				80.5		83.5		81.2		77.3

FeO* and Fe*, total iron as FeO and Fe, respectively. Mg# = Mg/(Mg + Fe²⁺) atomic ratio. Cr, Al, and Fe³⁺ = atomic fractions to Cr + Al + Fe³⁺ in spinel. (Fe), (Mg), and (Ca) = atomic fractions to Fe²⁺ + Mg + Ca. Cr# = Cr/(Cr + Al) atomic ratio. Fe²⁺ and Fe³⁺ were calculated assuming spinel stoichiometry. Fo = Fo content of coexisting olivine.

Fig. 9. Relationships between TiO₂ wt % and Fe³⁺/(Cr + Al + Fe³⁺) atomic ratio of spinels. Note the linear compositional trend for all spinels. Abbreviations are as described in Fig. 6. Arrow indicates an approximate variation outward in the orthopyroxene/basanite reaction product (see text).

Fig. 10. Relationships between Cr/(Cr+Al) atomic ratio and Mg/(Mg+Fe²⁺) atomic ratio of spinels in the Okete basanite and its peridotite xenoliths. Abbreviations are as described in Fig. 6.

hedral with only thin overgrowth of Fe-rich rim) to being almost perfectly euhedral (Fig. 2bc). Overgrowth is the most important process for the morphological modification of xenocrystal olivine in the Okete basanite. The overgrowing part sometimes distinctively inherits crystallographic orientation of the xenocrystal part and frequently includes minute spinels. The third is an “authentic” phenocryst in a narrow sense, which has been nucleated from the basanite magma due to inevitable olivine oversaturation during pressure release and cooling (Fig. 2a). The subhedral olivine with spinel inclusions around the xenolith rim (Fig. 4) is essentially equal to this type of phenocryst; both olivines were precipitated directly from the melt. The dimension of the subhedral olivine is sometimes constrained by the width of subgrains within olivine of xenoliths as mentioned above (Fig. 4a). It is difficult to distinguish the authentic phenocryst within the basanite matrix by observations with the optical microscope and back-scattered electron images. This type of olivine may be euhedral or skeletal or even dendritic highly dependent on the cooling rate (Donaldson, 1976).

2. Diversity of olivine-spinel pairs in the Okete basanite

All spinels as well as surrounding olivine phenocrysts were precipitated in chemically heterogeneous basanite magma, depending on the distance from peridotite xenoliths or peridotite-derived xenocrysts. The spinel in the orthopyroxene/melt reaction product, for example, were formed at high Cr# environments because mantle orthopyroxene has high Cr#s relative to ordinary magmas (e.g., Arai and Abe, 1995). The relatively high Cr# for chromian spinels in the orthopyroxene/melt reaction product is due to the high Cr# character of the secondary melt produced by the reaction. Crystallization of olivine and chromian spinel may have changed the relative amounts of Cr, Al and Fe³⁺ of the melt; the spinels were possibly changed from high-Cr# and low-Ti, Fe³⁺ to low-Cr# and high-Ti, Fe³⁺ with fractional crystallization of the basanite magma (cf., Arai, 1994b). The outward change of olivine and spinel compositions in the orthopyroxene/melt reaction product (Figs. 6 to 9) is ascribed to modification of olivine+spinel (= additional crystal growth and chemical change due to diffusion) by the surrounding basanite

magma. The olivine phenocryst (I) with spinel inclusions of the low- Fe^{3+} group possibly has an origin of the orthopyroxene/melt reaction, because the spinel inclusions are similar to those in the reaction product, especially in its outer part, both in the Cr-Al- Fe^{3+} ratio and TiO_2 content (Figs. 7 to 9). The former, however, is not exactly the same as the latter, the former spinel being, for example, a little more enriched in Fe^{3+} at a given TiO_2 content (Fig. 8). This is possibly due to higher diffusivity for Fe^{3+} in olivine than for Ti during the secondary modification from the basanite melt after the olivine was released from the xenolith boundary (e.g., Scowen *et al.*, 1991). The Fo content of olivine phenocryst (I) and Mg# of its low- Fe^{3+} group spinel inclusion are distinctly higher than the values for the reaction product (Figs. 6 and 10). This is due to Mg-Fe exchange with the evolved basanite magma, because diffusivity of Mg and Fe^{2+} in olivine is much higher than that of Al, Fe^{3+} , Cr and Ti (e.g., Scowen *et al.*, 1991).

The Fe-rich rim with spinel inclusions (width $\sim 50 \mu$) around the olivine phenocrysts (II) (Fig. 3) was formed by prevalent overgrowth on the olivine xenocryst from the basanite magma, because it is impossible to form the spinel-bearing rim of the olivine by only Fe-Mg exchange with the magma. The olivine xenocrysts were released to the basanite magma at the very late stage of evolution, which resulted in precipitation of Fe-rich olivine mantle with evolved (Ti-, Fe^{3+} -rich) spinel inclusions. The Mg-rich environment was, however, also available locally around the Mg-rich olivine phenocryst of xenocrystal origin (ca. Fo_{90}). It is noteworthy that some of the rim of the olivine phenocryst (II) is relatively magnesian (up to Fo_{85}) (Fig. 6) despite the evolved (Ti-, Fe^{3+} -rich) character of the spinel inclusions. It is highly possible that the olivine phenocrysts (I) with high- Fe^{3+} group spinels

which are similar in chemistry to those in the rim of the phenocrysts (II), were also precipitated from the basanite magma (Figs. 7 to 9). That is, the olivine phenocryst (I) with high- Fe^{3+} group spinel inclusion may be "authentic" phenocryst newly nucleated from the basanite magma.

The spinel inclusions in the subhedral part of olivine grains in the xenolith rim show relatively wide compositional ranges for the Cr#, Mg#, TiO_2 content and Fe^{3+} ratio, which almost cover the whole compositional ranges for spinels in the phenocrysts (I) and (II) and the orthopyroxene-basanite reaction product (Figs. 7 to 9). This wide variation of spinel composition is due to the chemical heterogeneity especially depending on the presence or absence of nearby pyroxenes along the xenolith/basanite boundary. Orthopyroxene, for example, can produce Si-, Cr-rich secondary melt on reaction. Alternatively or additionally the wide chemical variety of spinels means that the precipitation of olivine and its spinel inclusions had changed their compositions by fractional crystallization at the xenolith/basanite boundary from an early stage to a late stage of solidification of the Okete basanite.

3. Implications

Olivine phenocrysts in volcanic rocks, especially olivine-rich ones such as picrite basalts and some alkali olivine basalts, are possibly polygenetic. Primary magmas of deeper origin can be reactive with mantle peridotites, especially with their pyroxenes, at shallower part because they are successively oversaturated with olivine and undersaturated with pyroxene components with a decrease of pressure (e.g., Kushiro, 1969). This means that the reaction between peridotite and melt, which is essentially the same as that observed in the xenolith/basanite junction in the Okete basanite as well as in other xenolith-bearing magmas (Arai *et al.*, 1994), commonly occurs

within the upper mantle. This results in wide compositional ranges for olivine phenocrysts and their spinel inclusions in some olivine-rich volcanics such as the basanite from Okete Quarry, New Zealand (this study), and the picrite basalt from Oshima-Oshima volcano, the Northeast Japan arc (Ninomiya, 1993). It is concluded that the morphologically and chemically modified xenocrystal olivine, the mantle orthopyroxene/melt reaction product and the authentic phenocryst nucleated from the melt can become euhedral or nearly euhedral olivine phenocrysts within some magmas. If the morphological and chemical modification is very strong, olivine phenocrysts (with spinel inclusions) which have almost the same chemical characteristics but have various origins can be produced within high-temperature olivine-rich magmas.

VI. Summary and conclusions

1. The olivine phenocrysts in the Okete basanite are variable in morphology, texture and chemistry. The phenocrysts (I) are euhedral, strain-free and relatively small in size, and have spinel inclusions sparsely and evenly in the grain. The phenocrysts (II) are various in morphology (anhedral to euhedral), deformed or fractured (now healed) and relatively large in size, and have spinel inclusions only in the rim.

2. Mantle-derived peridotite (lherzolite) xenoliths are strongly reacted with the basanite melt at the boundaries, producing two kinds of olivines. One is the reaction product between orthopyroxene in the peridotite and the basanite melt. The olivine is subhedral to anhedral and has spinel inclusions. The other is subhedral olivine with spinel inclusions which has been overgrown on peridotite olivine. In addition to these the peridotite olivine is frequently released into the basanite matrix.

3. The spinels in the olivine phenocrysts generally show a compositional trend: the Cr#

is negatively correlated with Fe^{3+} and TiO_2 contents. The spinels are divided into two, low- Fe^{3+} group and high- Fe^{3+} groups. These parameters are slightly or not correlated with the Fo content of associated olivine as a whole. The olivine phenocrysts demonstrate the normal chemical zoning; from Fo_{85-77} to Fo_{77} outward for the phenocrysts (I) and from Fo_{90-91} (compositionally flat core) to Fo_{77} for the phenocrysts (II).

4. Mineral chemistry and petrographical characteristics suggest the various origins for the olivine phenocrysts. The olivine phenocrysts (I) with low- Fe^{3+} , high-Cr# spinels were derived from the orthopyroxene-basanite melt reaction product. Those with high- Fe^{3+} , low-Cr# spinels were precipitated directly from the basanite melt. The phenocrysts (II) are of xenocrystal origin but have been modified both chemically and morphologically to various degrees.

5. Olivine phenocrysts in olivine-rich volcanics such as picrite basalts and alkali olivine basalts are possibly polygenetic due to the peridotite/melt reaction and associated mechanical disintegration of mantle olivines in the peridotite.

Acknowledgements: We would like to deeply thank the late Professor Teruhiko Sameshima, Shizuoka University, for supplying one of us (S.A.) the samples used in this study. The samples were originally donated to Professor Sameshima by Professor R.M. Briggs, University of Waikato, New Zealand. We have been benefitted by discussions with A. Ishiwatari and A. Toramaru. Thoughtful reviews of two anonymous referees were helpful to improve the manuscript.

References

- Arai, S. (1987), An estimation of the least depleted spinel peridotites on the basis of olivine-spinel mantle array. *Neues Jahrb. Mineral. Monatsh.*,

- 347-357.
- Arai, S. (1994a), Characterization of spinel peridotites by olivine-spinel compositional relationships: Review and interpretation. *Chem. Geol.*, **113**, 191-204.
- Arai, S. (1994b), Compositional variation of olivine-chromian spinel in Mg-rich magmas as a guide to their residual spinel peridotites. *J. Volcanol. Geotherm. Res.*, **59**, 279-294.
- Arai, S. and Abe, N. (1995), Reaction of orthopyroxene in peridotite xenoliths with alkali basalt melt and its implication for genesis of alpine-type chromitite. *Amer. Mineral.*, **80**, 1041-1047.
- Arai, S., Abe, N. and Ninomiya, A. (1994), Reaction of peridotite xenoliths with host magmas as an analogue of mantle-melt interaction: microscopic characteristics. *Sci. Rep. Kanazawa Univ.*, **39**, 65-99.
- Boudier, F. (1991), Olivine xenocrysts in picritic magmas: An experimental and microstructural study. *Contrib. Mineral. Petrol.*, **109**, 114-123.
- Briggs, R.M. (1983), Distribution, form, and structural control of the Alexandra Volcanic Group, North Island, New Zealand. *New Zealand J. Geol. Geophys.*, **26**, 47-55.
- Briggs, R.M. and Goles, G.G. (1984), Petrological and trace element geochemical features of the Okete Volcanics, western North Island, New Zealand. *Contrib. Mineral. Petrol.*, **86**, 77-88.
- Donaldson, C.H. (1976), An experimental investigation of the delay in nucleation of olivine in mafic magmas. *Contrib. Mineral. Petrol.*, **69**, 21-32.
- Donaldson, C.H. (1978), Petrology of the uppermost upper mantle deduced from Calton Hill, Derbyshire. *Contrib. Mineral. Petrol.*, **65**, 363-377.
- Holmes, A. (1928), The Nomenclature of Petrology. pp.284, Thomas Murby & Co., London.
- Kelemen, P.B. (1990), Reaction between ultramafic rock and fractionating basaltic magma I. Phase relations, the origin of calc-alkali magma series, and the origin of discordant dunite. *J. Petrol.*, **31**, 51-98.
- Kushiro, I. (1969), The system forsterite-diopside-silica with or without water at high pressures. *Amer. J. Sci.*, **267-A**, 269-294.
- Ninomiya, A. (1993), Origin of the picrite basalt from Oshima-Oshima volcano, the Northeast Japan arc., Unpublished M. Sc. Thesis, Kanazawa University, Japan.
- Ninomiya, A. and Arai, S. (1993), Textural characteristics of olivine crystals in the Oshima-Oshima picrite basalt and their genetical implications. *Sci. Rep. Kanazawa Univ.*, **38**, 25-49.
- Robertson, D.J. (1976), A paleomagnetic study of volcanic rocks in the South Auckland area. Unpublished M. Sc. thesis, University of Auckland, Auckland, New Zealand. Quoted in Briggs and Goles (1984).
- Rodgers, K.A., Brothers, R.N. and Searle, E.J. (1975), Ultramafic nodules and their host rocks from Auckland, New Zealand. *Geol. Mag.*, **112**, 163-174.
- Sakuyama, M. (1978), Petrographic evidence of magma mixing in Shirouma-Oike volcano, Japan. *Bull. Volcanol.*, **41**, 501-512.
- Sakuyama, M. (1984), Magma mixing and magma plumbing systems in island arcs. *Bull. Volcanol.*, **47**, 685-703.
- Scowen, P.A.H., Roeder, P.L., and Helz, R.T. (1991) Reequilibration of chromite within Kilauea Iki lava lake, Hawaii. *Contrib. Mineral. Petrol.*, **107**, 8-20.

ニュージーランド、オケテ・ベイサナイト中の多起源性かんらん石斑晶

二ノ宮 淳・荒井 章司

ニュージーランド、オケテのベイサナイトおよびそのかんらん岩捕獲岩とメルトの反応物中のかんらん石、スピネルを記載した。かんらん石斑晶は二種類に大別できる。斑晶 (I) (コアで Fo_{85} - Fo_{76} , リムで Fo_{76}) は小型で自形性が高く、非変形で、細粒スピネルを包有する。斑晶 (II) (コアは Fo_{90} 以下で均質, リムは Fo_{85} - Fo_{76} の正累帯) は大型で、自形〜他形である。コアは変形組織を有し、リムのみスピネルを包有する。斑晶 (I) のうち、低 Fe^{3+} グループのスピネル (高 Cr #) を包有するものは、かんらん岩中の斜方輝石とメルトの反応物起源、高 Fe^{3+} グループのスピネル (低 Cr #) を有するものは、メルトから結晶化したものであろう。斑晶 (II) は外来結晶 (かんらん岩起源) 起源で、化学的、形態的に様々な改変を受けている。ピクライト玄武岩やアルカリかんらん石玄武岩のような、かんらん石斑晶に富む火山岩中のかんらん石斑晶は、メルトとかんらん岩との反応により、多起源性たり得る。