Melting of Amphibolite Xenolith Included in Peridotite of the Hidaka Metamorphic Belt, Hokkaido

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A fine-grained amphibolite xenolith was found in the Nikanbetsu peridotite body in the Hidaka metamorphic belt, Hokkaido, Japan. A thin crust of coarse-grained gabbro surrounds the xenolith cutting across the foliation of the amphibolite xenolith which is composed of kaersutite-pargasite, plagioclase, ilmenite, phlogopite and rutile. The gabbro crust consists of aluminous clinopyroxene, kaersutite-pargasite, plagioclase and ilmenite with minor hornblende, cummingtonite, phlogopite, olivine and spinel. The host rock is a plagioclase-spinel peridotite.

Major element chemical compositions of the amphibolite xenolith and the gabbro crust are similar to each other, suggesting that the gabbro crust is a melting product of the amphibolite xenolith. REE contents are also similar to each other, but higher in the gabbro crust than in the amphibolite xenolith. The bulk rock partition coefficients of REE between the xenolith and the gabbro melt suggest 35-40% melting for the production of the gabbro crust, whereas the structure and chemistry of the xenolith indicate a total melting of xenolith's rim, suggesting a migration of REE from the xenolith's interior to the melt.

The xenolith was included in a hot solid peridotite mass through solid intrusion of peridotites into the lower crust. During and after melting, Mg-Fe diffusion proceeded between the gabbro crust and the peridotites, and reached into about 1 cm depth of the peridotites. Based on the solid diffusion model on a polycrystalline olivine, 0.2 My is estimated for the duration of heating at minimum melting temperatures (about 900°C) of the gabbro. This rapid increase of temperature by the intrusion of hot solid peridotites may cause the lower part of the Main Zone to partially melt.

Introduction

The Hidaka belt of Hokkaido, Japan, is divided into two parts: the Western Zone and the Main Zone (Komatsu *et al.*, 1982). The former consists of meta-ophiolite (Miyashita, 1983), while the latter of various kinds of plutonic igneous and metamorphic rocks, rang-

ing from the greenschist to the granulite facies in grade (Osanai, 1985; Osanai *et al.*, 1986; Shiba, 1988). The Main zone includes also many kinds of peridotite bodies closely associated with metamorphics.

Migmatites occur widely in the granulite facies area (Osanai, 1985), between plutonic migmatites and metamorphics, and very rarely

Fig. 1. Generalized geological map of the Southmost Hidaka Mts.
A star indicates a locality of the xenolith.

between peridotites and metamorphics. Recently we found a gabbroic rock surrounding amphibolite included in a peridotite mass, which gives us a detailed information for the process of partial melting.

Outline of Geology

A generalized geologic map of the south-most Hidaka Mts. area is presented in Fig. 1. The eastern part of the area was mapped by Shiba (1988), and the western part by M. Tagiri, H. Suzuki, K. Suzuki, T. Inazawa and K. Tosaki of Ibaraki University. According to Shiba (1988), the lowest metamorphic grade zone is in the southeastern district near Shoya. The rocks were derived from mudstones and sandstones of the Mesozoic Nakanogawa Supergroup (Kontani, 1978). In the eastern part., the metamorphic grade increases northward, and reaches to the higher amphibolite facies in Zone D of Shiba (1988).

The metamorphic rocks along the Saruru-Toyoni-Furukawa shear zone (STF-STF) are strongly mylonitized by the upthrust of the Central Metamorphic and Igneous Composite Mass (CMIC Mass) which consists of high-grade metamorphics including granulites, migmatites, cordierite tonalite, gabbro-diorites and peridotites. At the time of upthrusting, the CMIC Mass was hot enough to make a thermal aureole in the zones A, B and C of Shiba (1988), though it is not recognized clearly in the southern marginal mylonite zone between the STF and the Horoizumi shear zone (H-H). Peridotite bodies and higher-grade metamorphics including granulites crop out frequently in the western part of the southern marginal mylonite zone. It is difficult to make a metamorphic zonal map in the southern marginal mylonite zone because of strong deformation and alteration.

Lower-grade metamorphic rocks of the CMIC Mass have similar mineral assemblages

to those of Zone D of Shiba (1988) outside of the mass. Cordierite tonalite are intruded into the central part of the metamorphic rock terrain. Furthermore, in many places pelitic and basic metamorphic rocks are migmatized (Fig. 1).

Fig. 2 shows the intrusion sequence of igneous rocks in this area. The rocks comprise peridotite, gabbro-diorite, tonalite and basalt. The peridotites have a remarkable layered structure consisting of dunite, spinel lherzolite, plagioclase lherzolite and gabbro (Niida, 1984). They are penetrated by gabbro-diorite dikes in the middle of the Rupeshupe River, and in a fault contact with mylonitic granulites at the Horoman reservoir. An amphibolite xenolith was found in the peridotite mass in the middle of Nikanbetsu River.

The gabbro-diorites are heterogeneous, and are divided into five types such as olivine gabbro, pyroxene gabbro, hornblende gabbro, hornblende diorite and cordierite bearing orthopyroxene diorite. The latter two include many metamorphic blocks and xenoliths, and in part, porphyritic plagioclase phenocrysts. The gabbro-diorites are frequently penetrated by

Fig. 2. The intrusion sequence of plutonic igneous rocks.

tonalites. The tonalites consist of cordierite tonalites and biotite tonalites. The former contain many metamorphic rock xenoliths, and are intruded so frequently into the metamorphic strata that migmatitic features are observed in many outcrops. Biotite tonalites are intruded into the cordierite tonalites (Shiba, 1988). After all, the basaltic dikes with a chilled margin are intruded into metamorphic and igneous rocks of the area.

Amphibolite Xenolith

An amphibolite xenolith was found in the peridotite body of the Nikanbetsu River. The host peridotite is a plagioclase (An=65)-spinel hercynite. The xenolith is an ellipsoidal disk, 20×10×3 cm in size. Its outline crosscuts the layering structure of the host peridotite (Fig. 3). Medium- to coarse-grained gabbroic crust, 3 to 5 mm in width, surrounds it (Fig. 4). The boundary between the xenolith and the crust is gradual, but that between the latter and the host peridotite is sharp (Fig. 4). Xenolith amphibolite is fine-grained, foliated and composed of kaersutite-pargasite, plagioclase (An=50), phlogopite and ilmenite with subordinate rutile. On the other hand, the gabbroic crust consists of aluminous clinopyroxene, kaersutite-pargasite, plagioclase (An=50), and ilmenite with minor hornblende, cummingtonite, phlogopite,

Fig. 3. Occurrence of the amphibolite xenolith. An arrow indicates the direction of layering in the host peridotite.

Fig. 4. Thin section photograph of the amphibolite xenolith. A=amphibolite, P=peridotite.

olivine and spinel.

Chemical compositions of the amphibolite, gabbro crust and peridotites are presented in Table 1. The analysed samples were obtained

Table 1. Chemical compositions of bulk rocks

wt%	Amphibolite	Gabbro	Peridotite-1	Peridotite-2
SiO ₂	41.58	40.99	42.03	40.74
TiO ₂	6.45	6.51	0.27	0.08
Al ₂ O ₃	16.99	13.99	2.83	2.50
Fe ₂ O ₃	1.93	1.69	2.44	3.17
FeO	8.86	9.14	10.88	5.32
MnO	0.14	0.15	0.21	0.13
MgO	9.30	12.52	34.33	36.77
CaO	9.51	10.65	2.53	1.60
Na ₂ O	3.26	2.17	0.44	0.17
K ₂ O	0.21	0.16	0.04	0.02
H ₂ O ⁺	1.69	1.93	3.40	8.55
H ₂ O ⁻	0.08	0.10	0.08	0.30
P ₂ O ₅	0.01	0.01	0.01	0.01
NiO	---	---	0.13	0.25
Cr ₂ O ₃	---	---	0.38	0.39
ppm				
Ba	111	88	23	38
Cu	112	207	42	13
Nb	25	24	5	3
Sc	36	58	13	9
Sr	726	883	53	44
V	611	556	61	40
Zn	76	84	110	68
Zr	38	82	18	25
La	1.9	2.2	0.0	0.0
Ce	7	9	1	1
Nd	5.6	9.2	0.3	0.0
Sm	1.27	2.91	0.0	0.0
Eu	1.31	1.55	0.18	0.09
Gd	2.4	4.4	0.7	0.7
Dy	2.42	4.49	0.52	0.20
Er	1.55	3.75	0.81	1.03
Yb	1.06	1.96	0.29	0.12
Lu	0.10	0.22	0.0	0.0
Y	11.25	20.7	2.7	1.3

by cutting the hand specimen to wafer-thin plates. Peridotite-1 is a portion of about 1 cm width in contact with the gabbro crust and Peridotite-2 lies outside Peridotite-1. The gabbro has a similar chemical composition to that of amphibolite. TiO_2 content of the amphibolite is enormously high. Among these rocks, SiO_2 , MgO and H_2O contents increase, and Al_2O_3 , Na_2O and K_2O contents decrease from the amphibolite towards Peridotite-2. In contrast to these components, total Fe content becomes maximum in Peridotite-1. CaO and TiO_2 contents of the amphibolite are similar to those of the gabbro crust, but Peridotites 1 and 2 are poor in them. The chemical change in these rocks would be explained by two models. One is the melting of the amphibolite xenolith, and the other the chemical interaction by diffusion between the gabbro crust and Peridotite-2.

Chemical compositions of amphiboles, pyroxenes, olivine, spinels and phlogopite were

determined by means of an EPMA, and presented in Table 2. Chemical compositions of Ca-amphiboles are plotted on the diagrams for the nomenclature of Ca-amphiboles by Leake (1978) (Fig. 5). Ca-amphiboles in the amphibolite xenolith and the gabbro crust show a zonal structure similar to each other, that is, the core is red-brown kaersutite, the mantle brown pargasite and the rim pale green hornblende. The kaersutite core exsolves many fine-grained needles of rutile along the crystal axes. Pargasites in the amphibolite and the gabbro are poorer in Mg than those in the host peridotite. Fig. 6 shows Ti versus Al contents of amphiboles from the xenolith and the Horoman ultramafic complex. Chemical compositions of the amphiboles in the xenolith range from those of crustal amphiboles to those of mantle amphiboles (Gamble and Kyle, 1987). The composition of amphibole core is plotted in the field of crustal amphiboles. On the other hand, those of amphibole rim lie in the field of

Table 2. Chemical compositions of representative minerals

	Amphibole			Cpx		Opx	Ol	Phl	Sp	
	1	2	3	4	5	6	7	8	9	10
SiO_2	40.50	42.24	41.47	48.15	49.70	54.53	38.54	38.24	0.03	0.05
TiO_2	6.01	2.42	4.77	2.00	0.78	0.17	n.d.	1.18	0.00	0.44
Al_2O_3	13.50	15.24	13.18	6.57	5.51	2.32	n.d.	19.49	64.46	34.41
Cr_2O_3	0.53	0.02	0.05	0.03	1.02	0.32	n.d.	0.02	0.00	27.91
FeO	8.38	10.93	11.58	6.98	5.56	9.14	17.58	9.57	23.93	28.21
MnO	0.12	0.06	0.11	0.18	0.18	0.28	0.30	0.08	0.15	0.26
NiO	0.00	0.05	0.07	0.08	0.00	0.00	0.29	0.17	0.13	0.22
MgO	15.35	12.73	12.40	13.26	15.07	31.54	43.14	18.98	12.93	8.18
CaO	11.29	11.73	11.45	21.17	21.07	0.93	0.00	0.06	0.02	0.05
Na_2O	2.34	2.49	2.26	0.71	0.73	0.04	n.d.	1.32	n.d.	n.d.
K_2O	0.68	0.45	0.61	0.00	0.00	0.00	n.d.	7.27	n.d.	n.d.
Total	98.70	98.36	97.95	99.13	99.62	99.27	99.85	96.38	101.65	99.73
Si	5.847	6.127	6.089	1.804	1.840	1.928	0.985	2.957	0.001	0.002
Al ^{IV}	2.153	1.873	1.911	0.196	0.160	0.072	n.d.	1.043		
Al ^{VI}	0.144	0.733	0.370	0.094	0.080	0.025	n.d.	0.732	1.970	1.223
Ti	0.652	0.264	0.527	0.056	0.022	0.005	n.d.	0.069	n.d.	0.010
Cr	0.060	0.002	0.006	0.001	0.030	0.009	n.d.	0.001	n.d.	0.665
Fe ³⁺	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.030	0.097
Fe ²⁺	1.012	1.326	1.422	0.219	0.172	0.270	0.376	0.619	0.489	0.615
Mn	0.015	0.007	0.014	0.006	0.006	0.008	0.006	0.005	0.003	0.007
Ni	n.d.	0.006	0.008	0.002	n.d.	n.d.	0.006	0.011	0.003	0.005
Mg	3.303	2.752	2.714	0.740	0.832	1.662	1.643	2.187	0.505	0.372
Ca	1.746	1.823	1.801	0.850	0.836	0.035	n.d.	0.005	0.001	0.002
Na	0.655	0.700	0.643	0.052	0.052	0.003	n.d.	0.198	n.d.	n.d.
K	0.125	0.083	0.114	n.d.	n.d.	n.d.	n.d.	0.717	n.d.	n.d.

2 in amphibolite, 1, 3, 4, 8 and 9 in gabbro, 5, 6, 7 and 10 in peridotite

Fig. 5. Chemical classification of the amphiboles.

Fig. 6. Ti vs. Al diagram for amphiboles. Data of the Horoman ultramafic complex are from Niida (1984). Fields of mantle and crustal amphiboles are from Gamble and Kyle (1987).

mantle amphiboles.

Fig. 7 shows the pyroxene compositions plotted on the quadrilateral diagram. Fig. 8 is the plot of Al^{VI} versus Al^{IV} for clinopyroxenes from the gabbro crust and the host peridotite. The diagram of $mg = Mg/(Mg + Fe)$ versus distance from the contact between the gabbro and the peridotite is given in Fig. 9. Clinopyroxenes in the gabbro crust are rich in Al_2O_3 (max. 6.5 wt%) and TiO_2 (max. 2 wt%), and exsolve orthopyroxene lamellae. Clinopyroxenes in the gabbro crust have chemical characteristics similar to those from xenoliths in basalts (Aoki and Shiba, 1973) (Fig. 8). Al, Ti and Fe contents of clinopyroxenes in the gabbro crust decrease toward the peridotite (Fig. 9). In the host peridotite, Mg contents of clinopyroxenes decrease toward the gabbro crust (Fig. 9). Clinopyroxenes in the peridotite are rich in Al_2O_3 (max. 5.9 wt%) and Cr_2O_3 (max. 1.0 wt%) and poor in TiO_2 (max. 0.7 wt%). CaO , Al_2O_3 and TiO_2 contents of orthopyroxenes in the host peridotite are less than 1.14 wt%, 3 wt% and 0.2 wt%, respectively. Mg contents of orthopyroxenes decrease toward the gabbro

Fig. 7. Quadrilateral diagram for pyroxenes and Fo contents of olivines. Abbreviations are the same as Fig. 5.

Fig. 8. Al^{VI} vs. Al^{IV} diagram for clinopyroxenes. Field boundaries are from Aoki and Shiba (1973).

crust (Fig. 9).

Fig. 7 and 9 show chemical compositions of olivines in the peridotite and the gabbro crust. Mg contents of olivines in the peridotite successively decrease toward the gabbro crust. An olivine in the gabbro crust is iron-rich ($mg = 0.74$). Chemical variation of spinels in the host peridotite is illustrated in Fig. 9. $Mg/Mg + Fe$ ratios of spinels also decrease toward the gabbro crust from the peridotite in concordance with the decreasing of Mg contents in olivines and pyroxenes. Concentration gradients of mg in constituent minerals of the peridotites and the gabbro crust would be explained by the diffusion between them.

Fig. 10 shows chemical compositions of phlogopites in the amphibolite xenolith and the gabbro crust. Plagioclases in the amphibolite and the gabbro crust do not show any apparent zonal structure.

REE Compositions

Rare earth element (REE) contents were analysed by means of an Inductively Coupled Argon Plasma Emission Spectrometer (ICP).

Fig. 9. Mg-Fe diffusion profiles in the gabbro and the peridotite. Tie-lines represent a single grain. $mg = Mg/(Mg + Fe)$.

Fig. 10. Chemical variation of phlogopite. Abbreviations are the same as Fig. 5.

The analytical procedure was described by Tagiri and Fujinawa (1988). REE compositions of the amphibolite, the gabbro crust, Peridotites -1 and -2 are presented in Table 1. Fig. 11 shows the REE normalized patterns on the basis of Leedy chondrite composition (Masuda *et al.*, 1973). REE contents of the gabbro crust are higher than those of the amphibolite xenolith. However, they are depleted in both light and heavy REE, so that the two REE patterns show similar curves convex upwards. The similarity of the REE pattern may exclude an opinion that the gabbro crust originated from an extraneous melt intruding into the boundary between the peridotite and the amphibolite xenolith. On the contrary, the similar REE patterns allow an idea that the gabbro crust is a melting product of the am-

Fig. 11. REE normalized patterns of the amphibolite and the gabbro crust.

phibolite xenolith.

Both the amphibolite xenolith and the gabbro crust show positive Eu anomaly, though a larger anomaly is observed in the former. It is suggested that kaersutite-pargasite selectively melted into the gabbro crust. This is consistent with the enrichment of heavier REE preferentially contained by hornblende in the gabbro crust. A trace of some REE, for example Eu, Dy and Yb, diffused into Peridotite-1 (Table 1).

Discussion

Partial Melting of the Amphibolite Xenolith

The mode of occurrence, major element chemistry and REE composition of the xenolith indicate that the amphibolite xenolith would have been melted partially or almost to form the gabbro crust, though melting textures are not observed in any crystal in the xenolith. Fig. 12 shows the bulk rock partition coefficients of major elements, REE and trace metals between the gabbro crust (=melt) and the amphibolite xenolith (=solid). The gabbro crust is depleted in Na, K and Al, and enriched with Mg, Ca and Sr, but Si, Ti, V, Mn, Fe and Nb are invariant. Heavy REE are highly concentrated in the gabbro crust except for Eu. The concentration of light REE is small in the

gabbro crust.

Here, we consider that melting occurred in a closed system, and that the xenolith was composed of solid and melt parts. Assuming the REE partition equilibrium between solid and melt, the following equation holds (Masuda, 1979),

$$\log(C/C_0) = (k-1) \log F_w \quad (1)$$

in which C_0 is the concentration of REE in the initial amphibolite xenolith, and C is the concentration of REE in melt at a partial melting degree (F_w). The coefficient, k is the bulk rock partition coefficient C'/C , where C' is the concentration of REE in solid. The k is assumed to be constant in the partial melting process. Though the boundary between the amphibolite xenolith and the gabbro crust is gradual, we presume that their chemical compositions represent those of a solid and a melt in the xenolith.

If the melt did not migrate into the peridotite, the initial volume of the xenolith would have been the sum of present volumes of the amphibolite xenolith and the gabbro crust. When we adopt an elliptic disk of $205 \times 105 \times 35$ mm for the initial volume of xenolith based on the mode of occurrence, we get the volumetric

Fig. 12. Bulk rock partition coefficients.

ratio 1:4 for the gabbro crust to the amphibolite xenolith. Using this volumetric ratio, we can estimate a concentration of REE in the initial amphibolite xenolith before melting. Substituting all data of REE for the equation (1), we get $Fw=0.34-0.43$.

On the other hand, the bulk rock partition coefficients of major elements are about 1 (see Fig. 12), and $Mg/(Mg+Fe)$ of the gabbro crust is very similar to that of the amphibolite, which indicates a complete melting of the xenolith. As any migmatitic structure is not observed in the amphibolite xenolith, it is hardly to explain ca. 40% partial melting of the xenolith. It is concluded that the gabbro crust was produced by whole melting of the rim of the xenolith and some REE migrated from the interior of the xenolith to the melt.

Osanai *et al.* (1986) estimated 730°C and 6.5 kb for the P-T conditions of the granulite facies in the Main Zone, and concluded that these values correspond to the physical conditions at 23 km in depth of the lower crust. The peridotite body would have been in a deeper

level. When we assume ca. 30 km in depth for the location of the peridotite, 960°C and 8.5 kb are given for the P-T conditions of the solid peridotite. These physical conditions are consistent with Niida's consideration (850–1000°C at Moho) (Niida, 1984). According to the experiments by Hariya *et al.* (1974), kaersutite and hornblende break down at 870°C and 920°C, 8.5 kb, respectively. An amphibolite including other minerals may begin to melt at a lower temperature, so that the amphibolite xenolith might receive enough heat to melt entirely.

Mg-Fe Diffusion between the Gabbro Crust and the Peridotite

Fig. 9 shows the Mg-Fe diffusion patterns of constituent minerals in the gabbro crust and the peridotite. In polycrystalline natural rocks, Mg-Fe transport may occur by volume diffusion and boundary diffusion (Manning, 1974). The interface of the present case is the boundary between the gabbro crust and the peridotite. When the diffusion coefficient is known, we can estimate the heating duration or

a mean path of transport based on the following equation ;

$$\text{Mean path of transport} = \sqrt{2Dt} \quad (2)$$

(Miyashiro, 1964).

Here, D and t are diffusion coefficient and time, respectively. Misener (1974) determined interdiffusion coefficients of Mg and Fe in olivine (Fo10-80) at 900-1100°C, and 1 atm-35 kb. From his experiments, we adopt an interdiffusion coefficient of Fo80, ca. $10E-15$ cm²/sec at 900°C and 10 kb. A mean path of transport in peridotite is supposed to be 1 cm from the interface, though it is less than 0.5 mm in the gabbro crust. Then, we obtain ca. 20 My for a heating duration of the xenolith at 900°C. A diffusion coefficient of Mg-Fe in spinel is ca. $10E-11$ cm²/sec at 1200°C and 1 atm (Lehmann, 1983). As it is still unknown at 900°C, 10 kb, we cannot estimate any heating duration from spinel composition.

The interdiffusion coefficient must be extremely small compared to the apparent diffusion coefficient in natural rocks, because the interdiffusion coefficient was determined in the diffusion experiment between single crystals. In the diffusion experiment of ¹⁸O in albite (Giletti *et al.*, 1978), the coefficient in powder material is larger by the four order of magnitude than that in a single crystal. In the grain boundary diffusion process, the diffusion coefficient in polycrystalline olivine may increase by the four order of magnitude. When we adopt the larger value of the coefficient, the heating duration decreases by the two order of magnitude. The heating duration at 900°C, 0.2 My, is very short for the intrusion of hot solid peridotite. For example, a cooling rate of the Glen Dessary syenite from ca. 1000°C to ca. 700°C was estimated to be ca. 30°/My by Cliff (1985). This rapid increase of temperature by the intrusion of the peridotite mass may cause the lower part of the Main

Zone to melt partially.

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北海道日高変成帯のかんらん岩体中に包有される 角閃岩捕獲岩の部分溶融

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日高変成帯ニカンベツかんらん岩体中に細粒の角閃岩捕獲岩を見出した。角閃岩はガブロの薄層に囲まれて、かんらん岩体の層構造を切るように捕獲されている。角閃岩はケルスート閃石-バーガス閃石+斜長石+イルメナイト、ガブロは単斜輝石+ケルスート閃石-バーガス閃石+斜長石+イルメナイトを主とする。両岩石は全岩化学組成が極めて似ている。REEパターンも両者で似ているが、ガブロの方がよりREEに富んでいる。これらの検討から、角閃岩はかんらん岩体に捕獲された時その縁が全溶融し、REEは捕獲岩の核部から移動したと考えた。ガブロとかんらん岩の間には顕著な拡散層があり、かんらん岩体中には約10 mmの深さまでFe, Mgの拡散が認められる。拡散係数を用いて900°Cでの加熱時間を推定すると、約0.2 Myとなる。この結果は、熱いかんらん岩体の進入によって、日高変成帯の下部が溶融したことを暗示する。