

Cooling process of a granitic body deduced from the extents of exsolution and deuteritic sub-solidus reactions: Case study of the Okueyama granitic body, Kyushu, Japan

Takashi Yuguchi*, Tadao Nishiyama

*Department of Earth and Environment, School of Science, Graduate School of Science and Technology,
Kumamoto University, 2-39-1, Kurokami, Kumamoto 860-8555, Japan*

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Abstract

The variation in cooling processes with depth in a magma body is evaluated quantitatively by analysis of the extent of exsolution coarsening and deuteritic coarsening as sub-solidus reactions. This method is applied to evaluation of the Okueyama granitic body of central Kyushu, Japan. Exsolution coarsening has produced microperthite textures in this body, while deuteritic coarsening has resulted in patchperthite, myrmekite, and reaction rims, respectively. Through measurement of six textural parameters, including the width and spacing of lamellae and the thickness of myrmekite and reaction rims, the extent of these sub-solidus reactions is shown to increase systematically with depth in the granite body, indicating that the Okueyama cooled gradually from the roof. The hornblende–plagioclase and ternary feldspar thermometers indicate temperature a range of 710 to 620 °C for volume diffusion associated with exsolution coarsening, while deuteritic coarsening is found to have occurred at temperatures below 500 °C on the basis of the ternary feldspar thermometer. The cooling period corresponding to exsolution coarsening is estimated using a one-dimensional heat transfer model, yielding periods of 820 y at the roof and 1390–1890 y at the base of the exposure (1000 m below the roof) depending on total depth of the original magma body.

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1. Introduction

The cooling of a magmatic body is an important process in crustal evolution, as the heat of a pluton causes contact metamorphism and may play a role in the development of regional low-*P/T* metamorphic belts (e.g. Hanson and Barton, 1989; De Yoreo et al., 1991). In terms of igneous processes, the study of magmatic

cooling contributes to determining the variation in the crystallization temperatures of plutonic rocks with depth.

Sub-solidus reactions in plutonic rocks are a potential explanatory tool for quantifying such cooling process. The two types of sub-solidus reactions that have been identified are exsolution and deuteritic reactions. Transmission electron microscopy (TEM) and HF-etching scanning electron microscopy (SEM) have revealed that K-feldspar phenocrysts in the Shap granite of Cumbria, northwest England, has undergone both an exsolution reaction to form microperthite with film lamellae and a deuteritic reaction to form patchperthite through

* Corresponding author. Tel./fax: +81 096 342 3411.

E-mail address: yugtk4@es.sci.kumamoto-u.ac.jp (T. Yuguchi).

dissolution–precipitation by deuteritic fluids (Worden et al., 1990; Waldron and Parsons, 1992; Waldron et al., 1994; Lee et al., 1995; Walker et al., 1995; Lee and Parsons, 1997). Hashimoto et al. (2005a) suggested that

an exsolution reaction in K-feldspar from the Tanakami granitic pluton of southwest Japan is responsible for the formation of microperthite with micrometer-scale lamellae and cryptoperthite with nanometer-scale

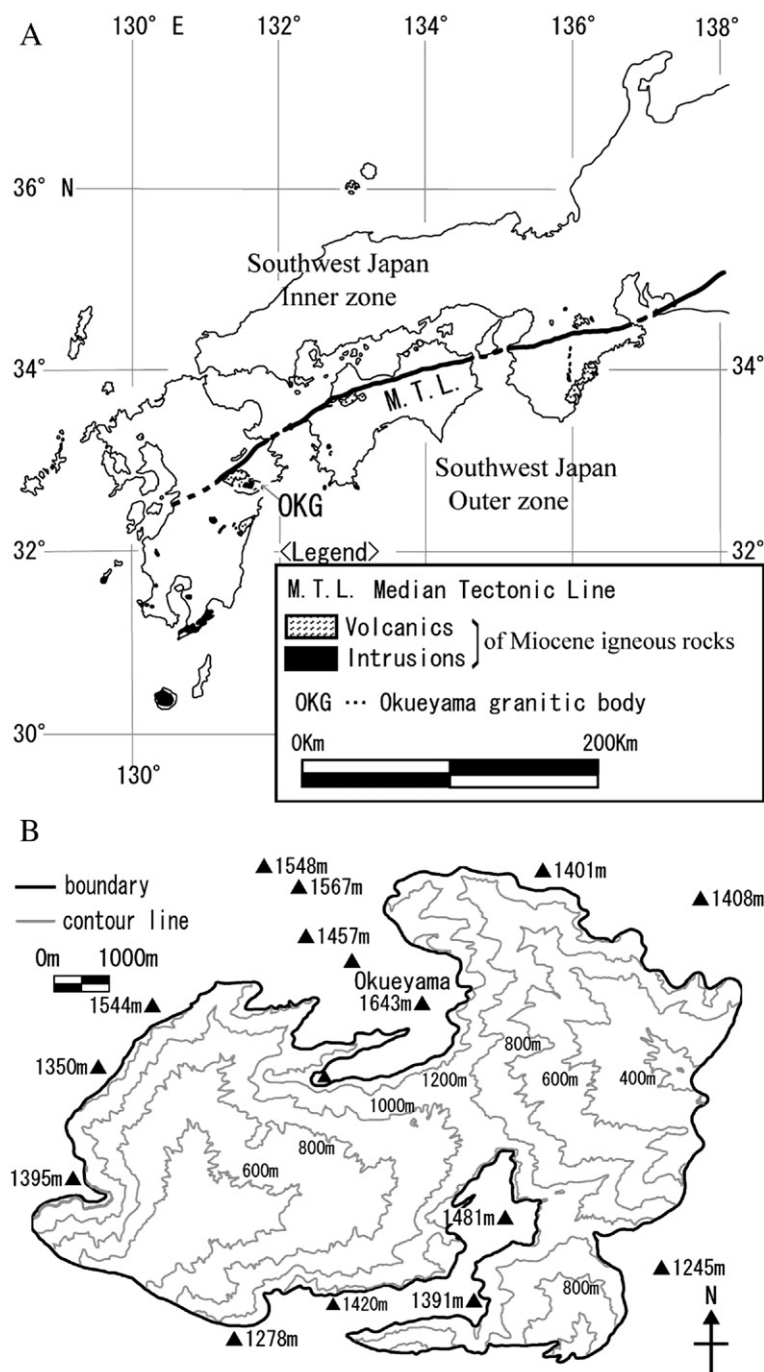


Fig. 1. The Okueyama granitic body. (A) Locality map showing the Okueyama granitic body (solid symbol) in Kyushu, and the distribution of felsic Miocene igneous rocks in southwest Japan (after Nakada and Takahashi, 1978). (B) Topography inside the Okueyama granitic body (contour interval, 200 m).

lamellae. Cryptoperthite and microperthite are formed by alkali inter-diffusion in K-feldspar (e.g., Brady and Yund, 1983; Christoffersen et al., 1983; Hokanson and Yund, 1986). Brown et al. (1983) examined the lamella periodicity in cryptoperthite from the Klokken gabbrosyenite of South Greenland by TEM, and found that the periodicity is composed of a long (first) and short (second) component. The lamella periodicity indicates that the spacing of cryptoperthite lamella in one K-feldspar grain is almost constant. Both lamellae periodicities become shorter with decreasing paleodepth, indicating that the rock body was effectively cooled from the shallower extent. The cooling time over the exsolution temperature range in the Klokken gabbrosyenite estimated by Brown and Parsons (1984) is several years at the top of the body and several hundreds to one thousand years at the bottom. The determination of Brown and Parsons (1984) is based on a combination of cryptoperthite lamella spacing and the Na–K inter-

diffusion coefficient obtained by the isothermal annealing experiments of Yund and Davidson (1978) and Yund et al. (1974).

In the present study, the extent of sub-solidus reactions recorded in the Okueyama granitic body of central Kyushu, Japan, is investigated as a basis for clarifying the cooling process and estimating the cooling period. The Okueyama granite represents a good example for such study in that the body consists of a single magma chamber, which solidified as a zoned pluton as described by Takahashi (1986). The Okueyama granite has very good vertical exposure of greater than 1000 m, corresponding to the full depth of the magma chamber.

Good indicators of the extent of sub-solidus reactions include albite lamella in microperthite and patchperthite, and the plagioclase/K-feldspar textures of myrmekite and a reaction rim. The development of these textures is examined quantitatively, and the variation of reaction extent is mapped according to altitude from 400 m

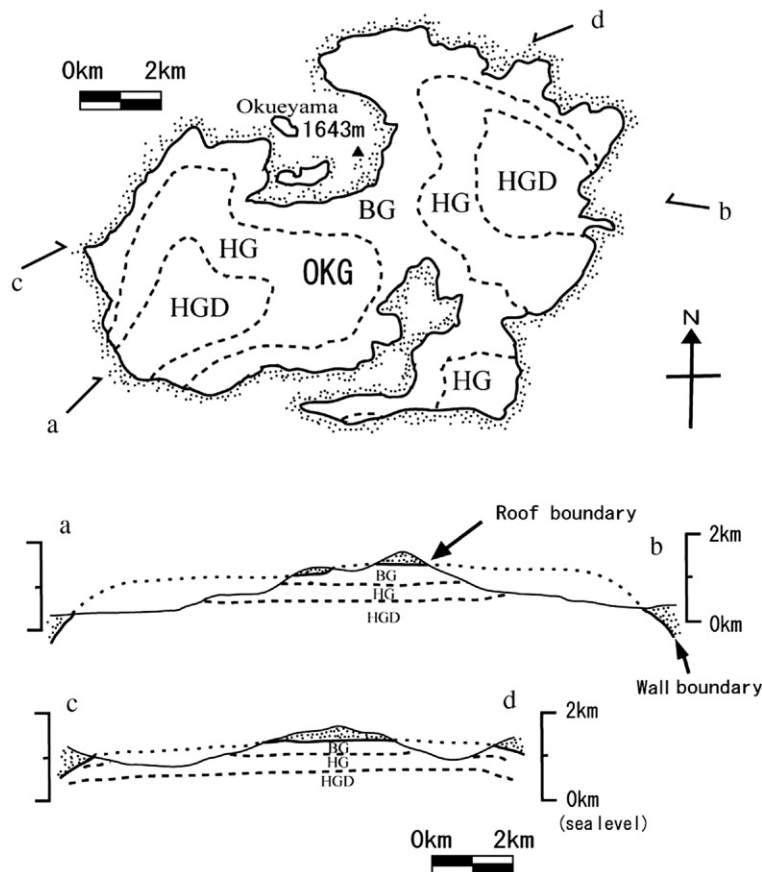
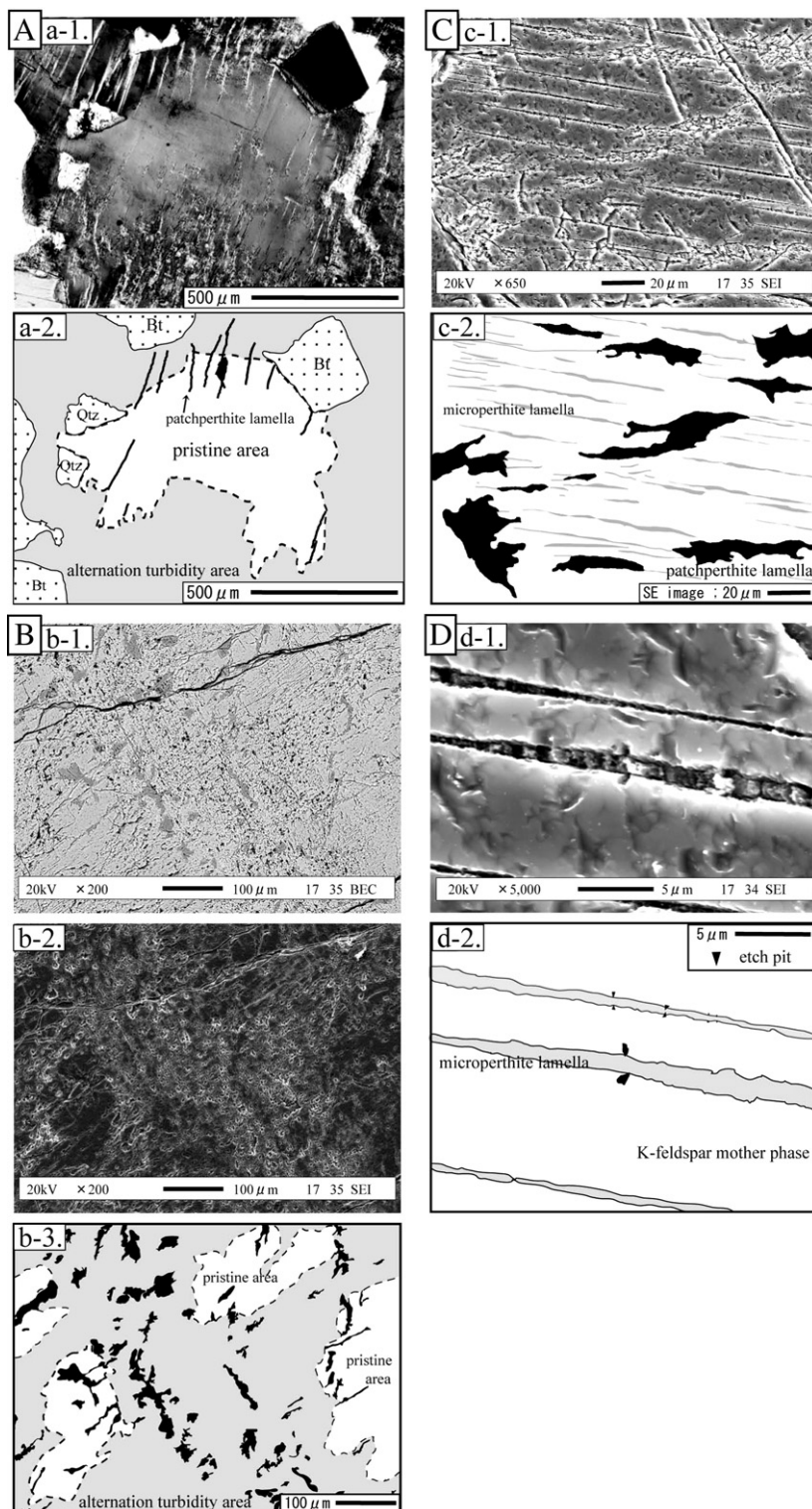


Fig. 2. Rock facies distribution and cross-section for the Okueyama granitic body (BG, biotite granite; HG, hornblende-biotite granite; HGD, hornblende-biotite granodiorite). "Reprinted from *Journal of Volcanology and Geothermal Research*, vol. 29, Masaki Takahashi, *Anatomy of a middle Miocene Valles-type caldera cluster: geology of the Okueyama volcano-plutonic complex, southwest Japan*, Page No. 33–70, Copyright (1986), with permission from Elsevier".

(bottom of exposure) to 1400 m (roof). Myrmekite is an intergrowth texture consisting of vermicular quartz and sodic plagioclase, while the reaction rim texture is

present as an albite-rich rim free of vermicular quartz. This reaction rim only occurs between plagioclase and K-feldspar. Myrmekite is relatively rare in the granitic



body, whereas the reaction rim is found throughout. These micrometer-scale textures can be observed by polarizing optical microscopy (POM) and SEM.

Six parameters are evaluated in order to determine the extent of sub-solidus reactions: the width and spacing of albite lamella in micropertthite and patchperthite, and the thickness of myrmekite and reaction rims. The hornblende–plagioclase and ternary feldspar thermometers are used effectively for determination of the temperatures of the two sub-solidus reactions (exsolution coarsening and deuteric coarsening). A one-dimensional (1D) heat-transfer model is used to estimate the period over which the sub-solidus coarsening reactions took place.

2. Geological setting

The Japanese Islands are divided into southwest and northeast sections by the Itoigawa–Shizuoka Tectonic Line. In southwest Japan, the Median Tectonic Line (MTL) demarcates the inner and outer zones. Felsic igneous rocks of Miocene age are distributed in Setouchi and the outer zone of southwest Japan, forming a belt 150 km wide and 800 km long, oriented parallel to the Nankai Trough (Nakada and Takahashi, 1979). This igneous activity in the outer zone is recent, occurring over a short period between 17 and 10 Ma with a peak near 14 Ma based on biotite and whole-rock K–Ar ages determined by Nakada and Takahashi (1979) and Nakada and Okamoto (1984).

The Okueyama granite is one member of the Miocene plutonics of the outer zone, and has been dated at 14 Ma (biotite K–Ar, Shibata, 1978; whole-rock Rb–Sr, Shibata and Ishihara, 1979). The body is located approximately 20 km south of the MTL in north Miyazaki, central Kyushu (Fig. 1). The Okueyama granite intrudes the Cretaceous Lower Shimanto Group (dated by radiolarian fossils) in the outer zone of southwest Japan. The Lower Shimanto Group has undergone low-grade regional metamorphism of prehnite–actinolite facies associated with subduction of the young and hot slab (Miyazaki and Okumura, 2002). The Okueyama granite is the largest ($9 \times 11 \text{ km}^2$ surficial extent) batholithic pluton among several isolated members forming the

Okueyama granitic complex (Takahashi, 1986), and its formation represents the final episode of magmatic activity in the Okueyama volcano–plutonic complex (Takahashi, 1986). The Okueyama granite has a flat roof boundary and a steeply dipping wall boundary (Takahashi, 1986, Fig. 2a), and contact metamorphism extending into the surrounding sediments grades from biotite through cordierite and orthopyroxene zones (Miyazaki and Okumura, 2002).

3. Petrography

3.1. Rock facies, mode, and bulk chemical composition

The Okueyama granite is characterized by vertical changes in rock facies, mode, and bulk chemical composition, interpreted as the result of gravitational fractionation of crystals in the magma chamber (Takahashi, 1986, 1987). The rock facies grades downward from biotite granite through hornblende–biotite granite to hornblende–biotite granodiorite (Fig. 2). K-feldspar and quartz decrease monotonically downward, whereas plagioclase and mafic minerals (hornblende and biotite) increase in the same direction (Takahashi, 1986). Hornblende does not occur in the upper part of the Okueyama granite (above 1070 m altitude). SiO_2 , K_2O and $\text{K}_2\text{O}/\text{Na}_2\text{O}$ decrease monotonically downward from the roof, whereas the other major oxides increase in the same direction (Takahashi, 1986).

3.2. Reaction textures during two sub-solidus reaction stages

The sub-solidus reaction textures were analyzed in order to clarify the cooling process in the Okueyama granite. Reaction textures such as perthitic K-feldspar, myrmekite, and the albitic reaction rim were examined under optical microscope. Minerals and textures were further observed using an SEM energy-dispersive spectroscopy (EDS) system (PC-SEM 5600, JEOL; combined with LINK ISIS) at Kumamoto University (acceleration voltage, 20 kV; beam current, 0.6 nA; beam width, 2 μm).

Fig. 3. Perthitic microtextures in K-feldspar. (A) POM image of turbid alteration and pristine zones in K-feldspar (sample No. 02091417). (a-1) Photomicrograph and (a-2) sketch of K-feldspar. Solid fragments in (a-2) indicate patchperthite lamellae intruding and cutting pristine zone. (B) BSE (b-1) and SE (b-2) image of turbid alteration and pristine zones in an HF-etched K-feldspar sample (No. 02051712). (b-3) Sketch showing intrusion of patchperthite lamellae (solid fragments) into pristine zone. (C) SE image (c-1) and sketch (c-2) showing the relationship between patchperthite lamellae and micropertthite lamellae in an HF-etched sample (No. 02091417). The periphery of the patchperthite lamellae hosts many micropores, corresponding to the turbid alteration zone observed by POM. Heavy patchperthite lamellae cut across thready micropertthite lamellae. (D) SE image (d-1) and sketch (d-2) of etch pits located at the interface between a micropertthite lamella and the K-feldspar mother phase in an HF-etched sample (No. 02091417). Pairs of etch pits occur at the interface as triangular wedges with edges turned inward toward the micropertthite lamella.

3.2.1. Microperthite and patchperthite

K-feldspar exhibits patchperthite lamellae, “turbid” alteration zones, and pristine zones (Fig. 3(A)), which are clearly visible by POM. The turbid zone, attributable to deuteritic alteration, is characteristically located surrounding the patchperthite lamellae, while the patchperthite lamellae intrude or cut the pristine area. Fig. 3(B) shows a backscattered electron (BSE) image and secondary electron (SE) image of the turbid and pristine zones in K-feldspar thin sections, which were prepared by etching in an ammonium hydrogen fluoride vapor by the technique of Waldron et al. (1994) followed by carbon coating. HF etching selectively etches the physically and chemically unstable areas, leaving concave defects such as micropores and etch pits. This process causes the albite lamella to stand out in bold relief from the orthoclase-rich phase, which is more readily etched, in the perthitic K-feldspar. After Hf etching, the turbid alteration zones become easily visible around the patchperthite lamellae, and present a high density of corrosion micropores, which are not observed in the pristine zone (Fig. 3(B,C)). The turbid zone is a common feature of K-feldspar, and is due to light scattering by micrometer-scale micropores (Worden et al., 1990). Patchperthite and micropores are formed by dissolution and reprecipitation, initiating at dislocation pits, during hydrothermal alteration accompanying a decrease in the volume of K-feldspar (Waldron et al., 1994). Patchperthite lamellae have low anorthite content (mean 6%, see Fig. 13(B)), even though K-feldspar of the Okueyama granite does not contain any anorthite. The spatial distribution of patchperthite and micropores, and the anorthite content in patchperthite lamellae support the notion that the patchperthite of the Okueyama granite was formed not by exsolution but by deuteritic alteration associated with hydrothermal influx from outside the K-feldspar grains.

The BSE and SE images (Fig. 3(B)) both indicate that many microperthite lamellae develop in the pristine zone. The front of the turbid zone and patchperthite lamellae intrudes and cuts the pristine zone, which contains microperthite lamellae. The microperthite lamellae are more thready than the patchperthite lamellae, and the interface between the albite lamellae and the K-feldspar mother phase in microperthite is sharp (semi-coherent), whereas the interface in patchperthite is irregular (incoherent; Lee et al., 1995). The albite-rich phases of patchperthite possess either a bead form or a lamella form. Fig. 3(C) shows that the orientation of microperthite lamellae is constant, whereas the orientation of patchperthite lamellae is variable. The heavy patchperthite lamellae cut across the thready micro-

perthite lamellae. Pairs of etch pits are located at the interface between the microperthite lamella and K-feldspar mother phase, forming triangular wedge shapes with edges pointing toward the microperthite lamella (Fig. 3(D)). The microperthite lamellae (including the etch pit area) may act as a pathway for deuteritic fluid and thus facilitate deuteritic alteration (Lee et al., 1995).

These observations imply that microperthite in the pristine zone is little affected by deuteritic alteration, indicating exsolution coarsening, whereas deuteritic alteration is the major cause of patchperthite formation.

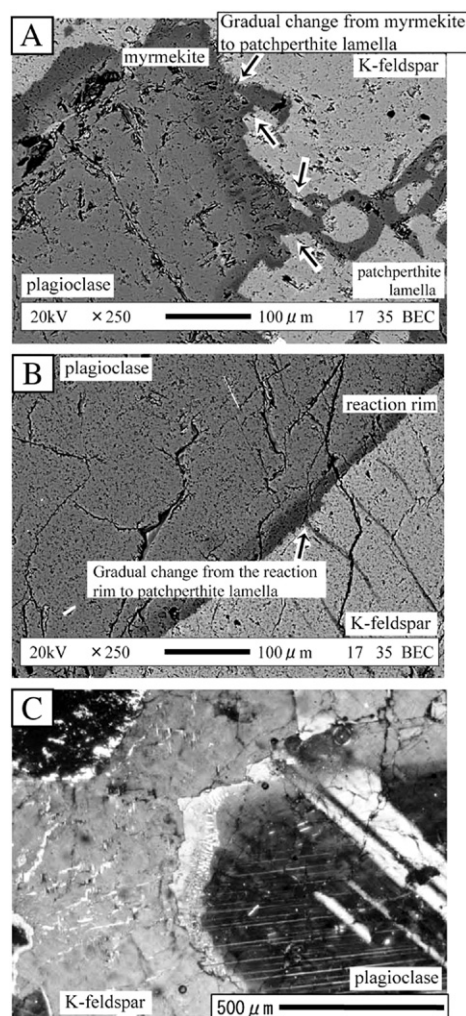


Fig. 4. Photograph showing the relationships among myrmekite, reaction rims, and patchperthite. (A) Gradual change from myrmekite to patchperthite lamellae (No. 02041830), (B) change from reaction rim to patchperthite lamellae (No. 02071415), and (C) composite texture consisting of myrmekite and reaction rim. Vermicular quartz (myrmekite) extends from plagioclase, but does not extend to K-feldspar. Myrmekite margin in contact with K-feldspar is free of quartz. No vermicular quartz is present in the reaction rim.

The effective conditions for the formation of these textures correspond to the exsolution coarsening stage (ECS) and the deuteric coarsening stage (DCS), respectively, where the present observations suggest that DCS occurred later than ECS.

3.2.2. Myrmekite and reaction rims

Myrmekite is an intergrowth texture consisting of vermicular quartz and Na-rich plagioclase that develops between K-feldspar and plagioclase (Smith and Brown, 1988). The reaction rim is defined here as an albite-rich rim, free of vermicular quartz, that develops at the contact between plagioclase and K-feldspar. As plagioclase in contact with other minerals does not show such rapid compositional and textural change, myrmekite and the albite-rich reaction rims are considered to form at a sub-solidus stage. For example, plagioclase in contact with

quartz exhibits a gradual compositional variation from $\text{Ab}_{59}\text{An}_{39}\text{Or}_2$ at the core (Ab, albite; An, anorthite; Or, orthoclase) to $\text{Ab}_{70}\text{An}_{29}\text{Or}_1$ at the rim. Myrmekite and the reaction rim are considered to form contemporaneously with patchperthite during the DCS, as indicated by the gradual change from myrmekite or the reaction rim to patchperthite lamellae (Fig. 3(A,B)). Similarly, there is a close relationship between myrmekite and the reaction rims as shown by a particular example of a composite texture consisting of both zones (Fig. 3(C)).

3.3. Extent of sub-solidus reactions

Each of the six parameters representing the extent of sub-solidus reactions in the two stages (ECS and DCS), the mean width and spacing of albite lamellae in microperthite and patchperthite, and the mean thickness of

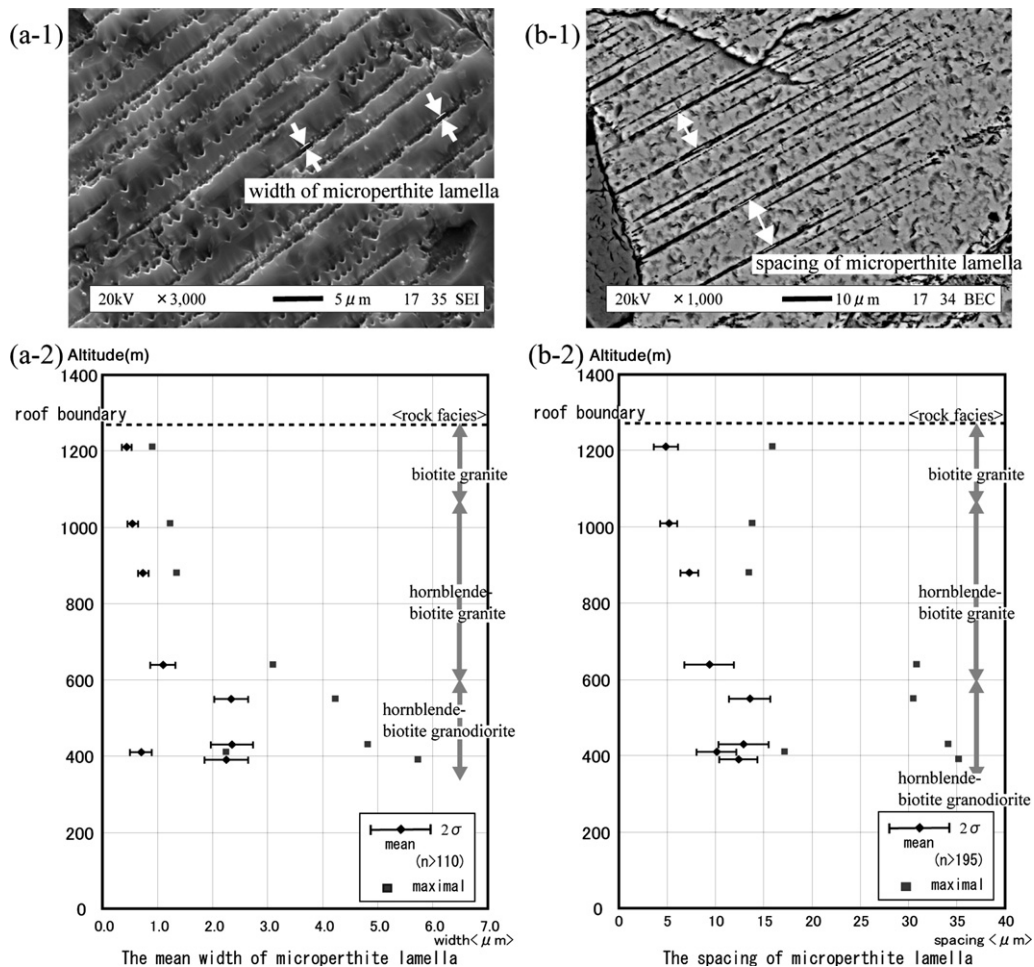


Fig. 5. Mean width and spacing of albite lamellae in microperthite. (a-1) SE image of microperthite lamellae in HF-etched sample (No. 02051712, altitude 410 m). Arrow indicates example of measured width. (a-2) Plot of mean width of microperthite lamellae with altitude. (b-1) BSE image of microperthite lamellae in HF-etched sample (No. 02091417, altitude 640 m). The arrow indicates an example of a measured spacing of microperthite lamella. (b-2) Plot of mean spacing of microperthite lamellae with altitude.

myrmekite and reaction rims, are detailed below. Microperthite textures represent ECS reactions, while patchperthite, myrmekite and reaction rim textures are the result of DCS sub-solidus reactions (Fig. 4).

3.3.1. Mean width of albite lamellae in microperthite

The mean width of albite lamellae in microperthite was determined based on more than 110 mea-

surements for each sample. Each measurement was taken as the widest part of a microperthite lamella (Fig. 5(a-1)). The mean width of albite lamellae in microperthite increases monotonically with paleo-depth in the magma body from 0.5 μm at the roof boundary to 2.5 μm at the lowest exposure (1000 m below the roof, corresponding to a modern altitude of 350 m, Fig. 5(a-2)).

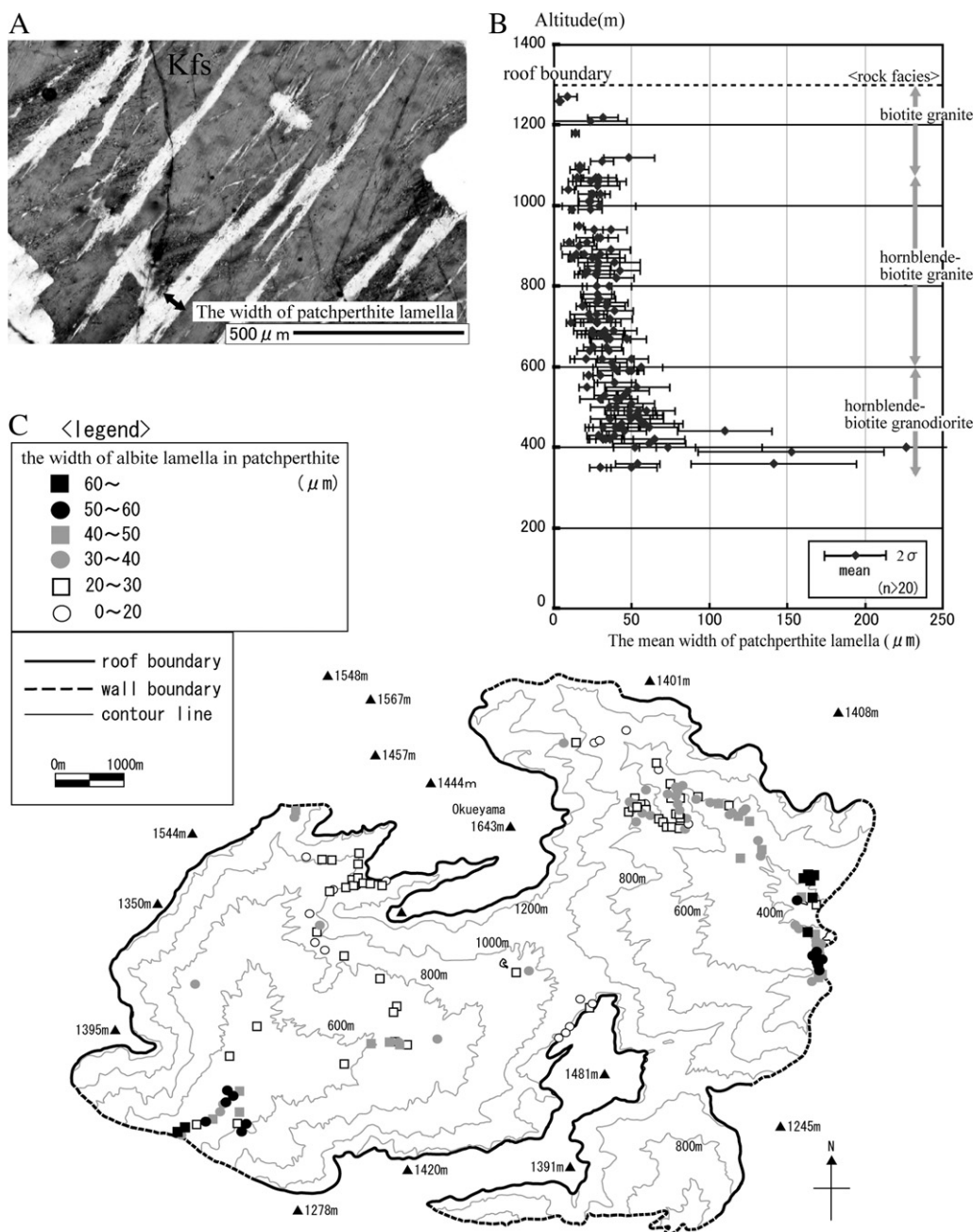


Fig. 6. Mean width of albite lamellae in patchperthite. (A) POM image of albite lamellae in patchperthite. Arrow indicates the widest part used for measurement. (B) Plot of mean width of patchperthite lamellae with altitude. (C) Plot of mean width of patchperthite lamellae on a topographic map.

3.3.2. Mean spacing of albite lamellae in microperthite

The mean spacing between neighboring albite lamellae in microperthite was determined based on more than 195 measurements for each sample. The spacing is defined as the distance between the center of one lamella and that of the neighboring lamella (Fig. 5(b-1)). The mean spacing of microperthite lamellae increases monotonically from

5 μm at the roof boundary to 15 μm 1000 m below the roof (Fig. 5(b-2)).

3.3.3. Mean width of albite lamellae in patchperthite

The mean width of albite lamellae in patchperthite was determined for 147 samples based on more than 20 measurements in each sample. Each measurement represents

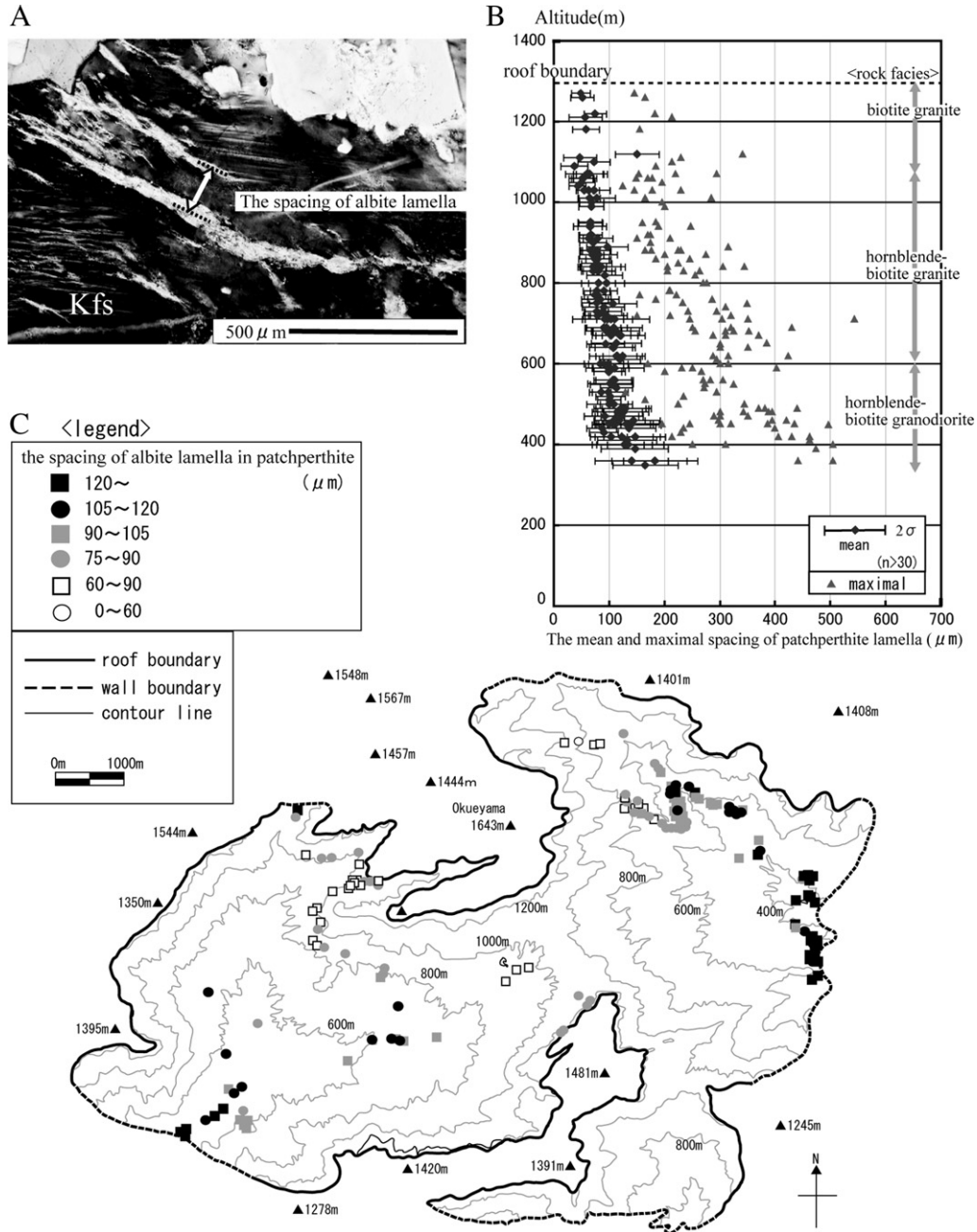


Fig. 7. Mean spacing of albite lamellae in patchperthite. (A) Photomicrograph of albite lamellae. Arrow indicates distance between centers of neighboring lamellae used for spacing measurement. (B) Plots of mean and maximum spacing of patchperthite lamellae with altitude. (C) Plot of mean spacing on a topographic map.

the thickest part of a lamella (Fig. 6(A)). The mean width increases monotonically from 10 μm at the roof boundary to 60 μm 1000 m below the roof (Fig. 6(B)). Fig. 6(C) shows a plot of the data on a topographic map. No systematic change from the wall boundary in the horizontal direction can be observed.

3.3.4. Mean spacing of albite lamellae in patchperthite

The mean spacing of albite lamellae in patchperthite was determined based on more than 30 measurements for each specimen (Fig. 7(A)). Fig. 7(B) shows a plot of the data against altitude for 146 samples, revealing a gradual downward increase from 50 μm at the roof to

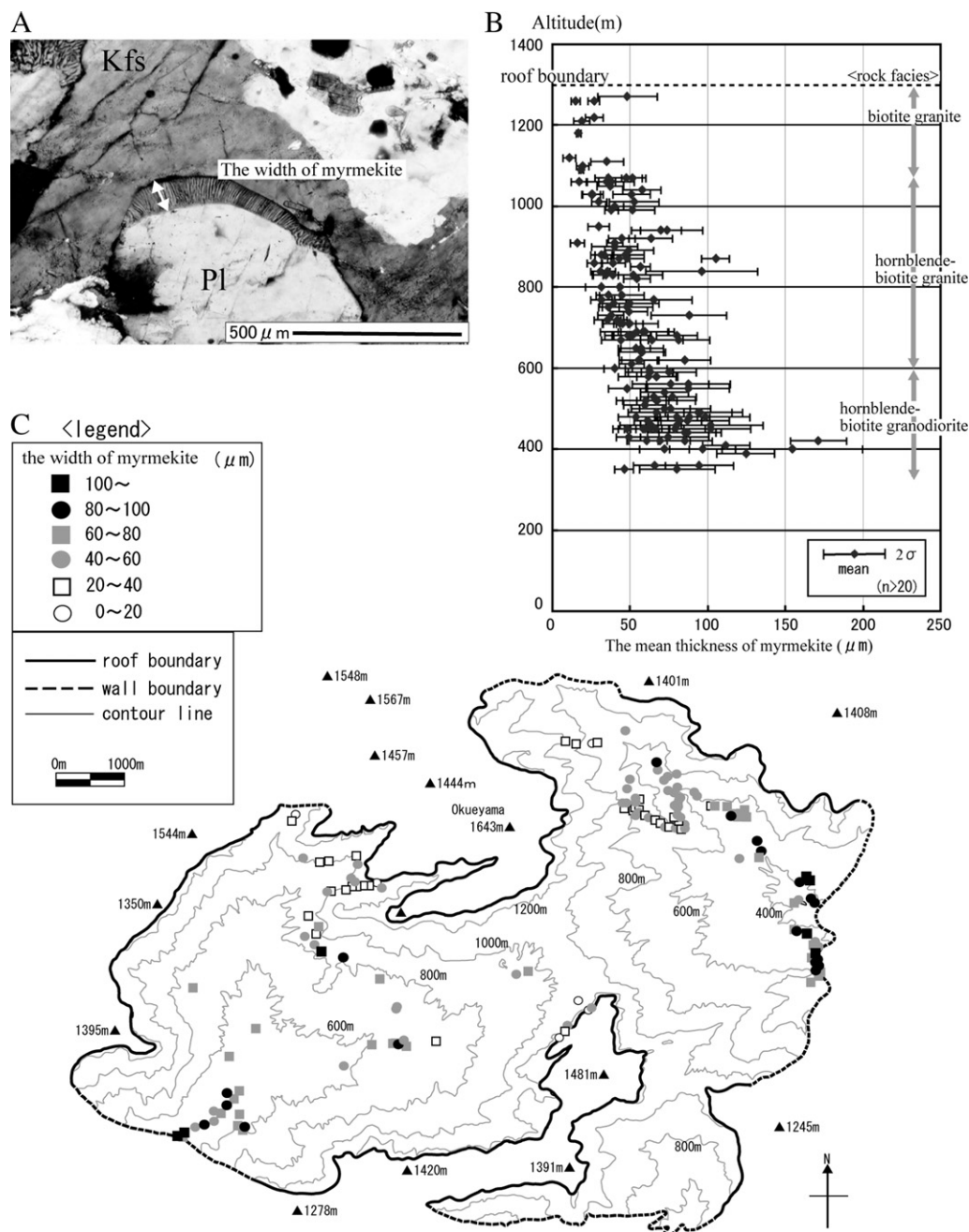


Fig. 8. Mean thickness of myrmekite. (A) POM image of myrmekite. Arrow indicates example of width used for measurement. (B) Plot of mean width of myrmekite with altitude. (C) Plot of mean thickness of myrmekite on a topographic map.

180 μm at the base of the exposure (1000 m below the roof). The maximum spacing also increases downward from 120 μm at the roof to 500 μm at the base of the exposure. No systematic horizontal variation can be observed in the mean spacing of patchperthite lamellae (Fig. 7(C)).

3.3.5. Mean thickness of myrmekite

The mean thickness of myrmekite was determined for 147 samples based on more than 20 measurements for each sample. The individual measurement represents the thickest part of a myrmekite zone (Fig. 8(A)). The mean width of myrmekite increases

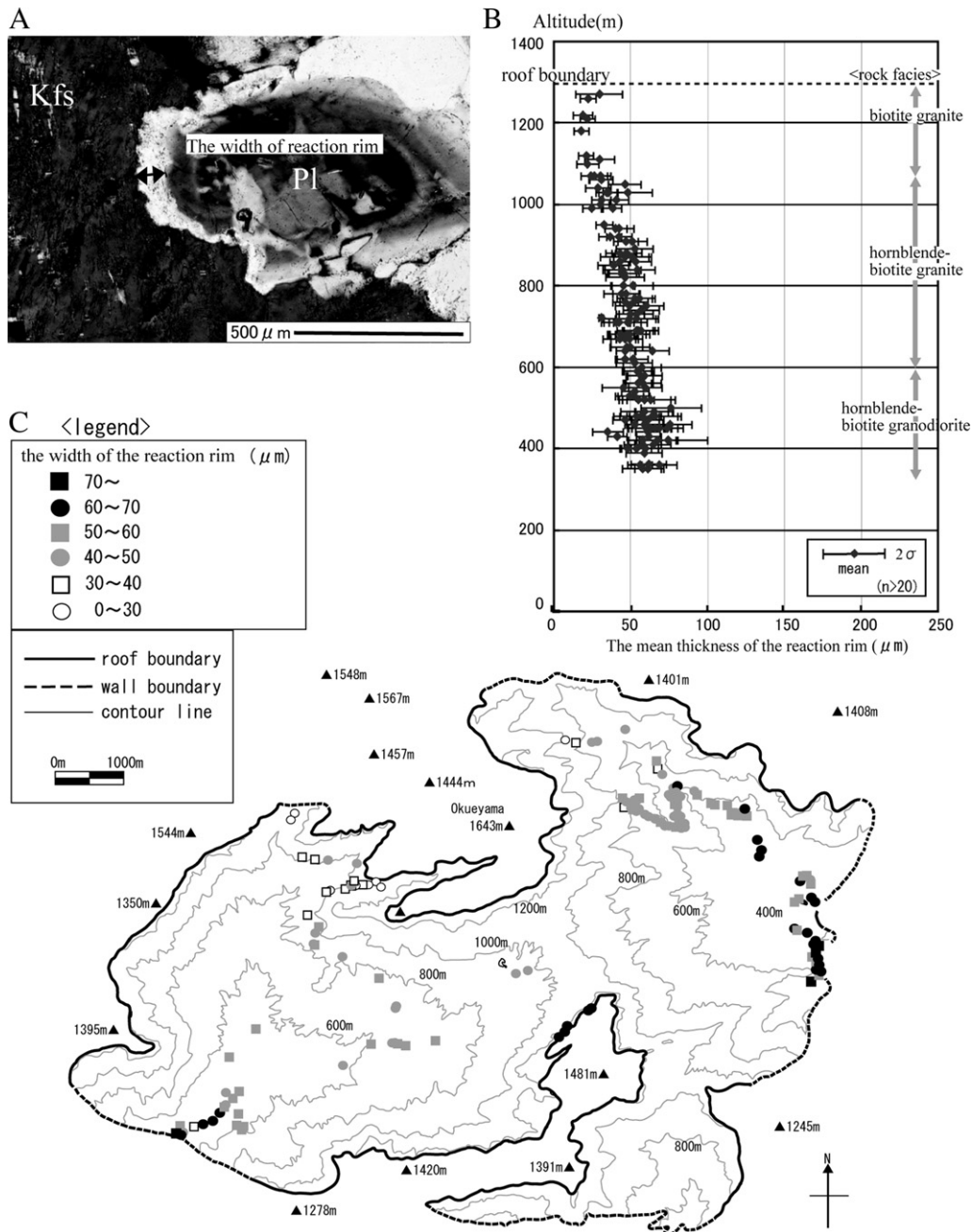


Fig. 9. Mean thickness of reaction rims. (A) POM image of reaction rim. Arrow indicates the thickest part used for measurement. (B) Plot of mean width of reaction rims with altitude. (C) Plot of mean width of reaction rims on a topographic map.

systematically from 10 μm at the roof boundary to 100 μm at the base of the exposure (Fig. 8(B)). A plot of the data on the topographic map (Fig. 8(C)) indicates no systematic horizontal variation from the wall boundary.

3.3.6. Mean thickness of reaction rims

The mean width of reaction rims was determined for 147 samples based on more than 20 measurements for each sample. Each measurement represents the thickest part of a reaction rim between plagioclase

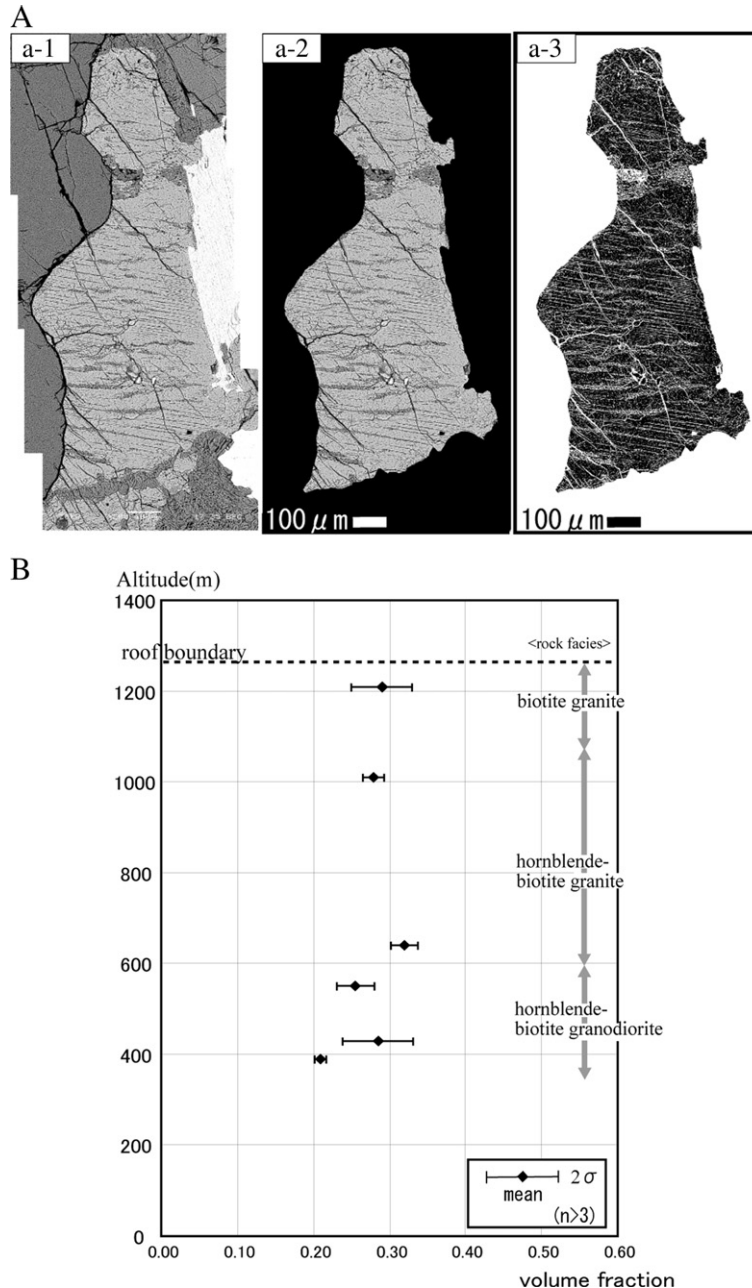


Fig. 10. Volume fraction of albite-rich phase in perthitic K-feldspar. (A) Volume fraction estimated from areal fraction of albite-rich phase in perthite grains. (a-1) Original image of a perthitic K-feldspar grain (No. 02091417, altitude 640 m). (a-2) Image with other minerals outside the K-feldspar grain removed. (a-3) Albite-rich phase and host orthoclase-rich phase shown as a binary image (B) Plot of volume fraction of albite-rich phase with altitude.

and K-feldspar (Fig. 9(A)). The values increase monotonically from 10 μm at the roof boundary to 80 μm at the base of the exposure (Fig. 9(B)). Fig. 9(C) shows a

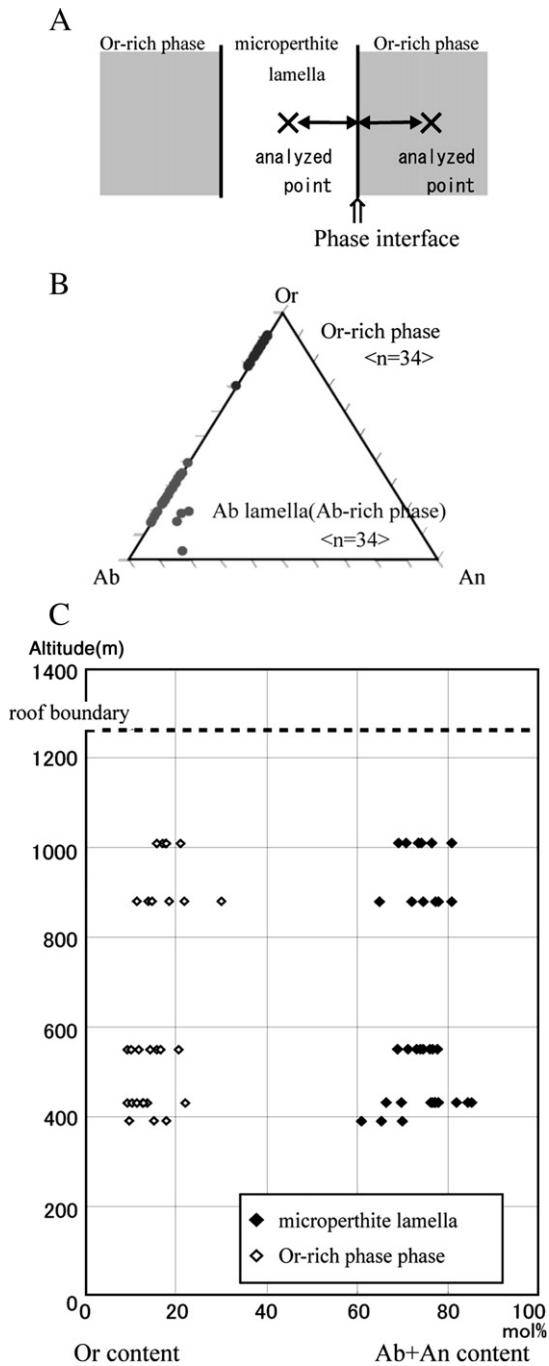


Fig. 11. Composition of albite lamellae and orthoclase-rich phase in microperthite. (A) Schematic figure showing set of analyzed points equally distant from the interface. (B) Plots of albite lamellae and orthoclase-rich phase compositions on an Or–Ab–An diagram. (C) Plots of albite lamellae composition and orthoclase-rich phase compositions in microperthite in terms of Or and Ab+An content against altitude.

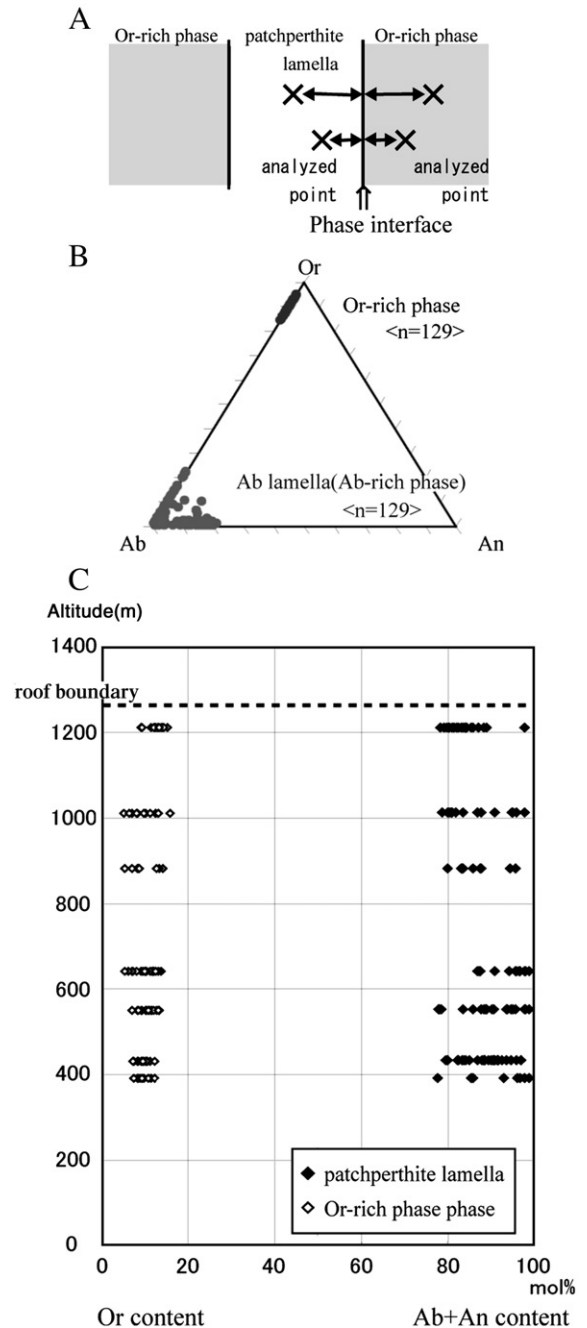


Fig. 12. Composition of albite lamellae and orthoclase-rich phase in patchperthite. (A) Schematic figure showing set of analyzed points equally distant from the interface. (B) Plots of albite lamellae and orthoclase-rich phase compositions on an Or–Ab–An diagram. (C) Plots of albite lamellae composition and orthoclase-rich phase composition in patchperthite in terms of Or and Ab+An content against altitude.

plot of the data on a topographical map, indicating no systematic change in the horizontal direction from the wall boundary.

3.4. Chemical composition of feldspar including sub-solidus reaction textures

The extent of sub-solidus reactions described above may depend on the duration of cooling and the bulk composition of feldspar. The change in feldspar composition with depth in the magma chamber, including the

sub-solidus reaction textures, was determined by measuring the volume fraction of the albite-rich phase in perthitic K-feldspar, the composition of albite lamellae and the orthoclase-rich phase in microperthite and patchperthite, the local bulk composition of K-feldspar grains, and the composition of myrmekite and reaction rims and the surrounding mineral.

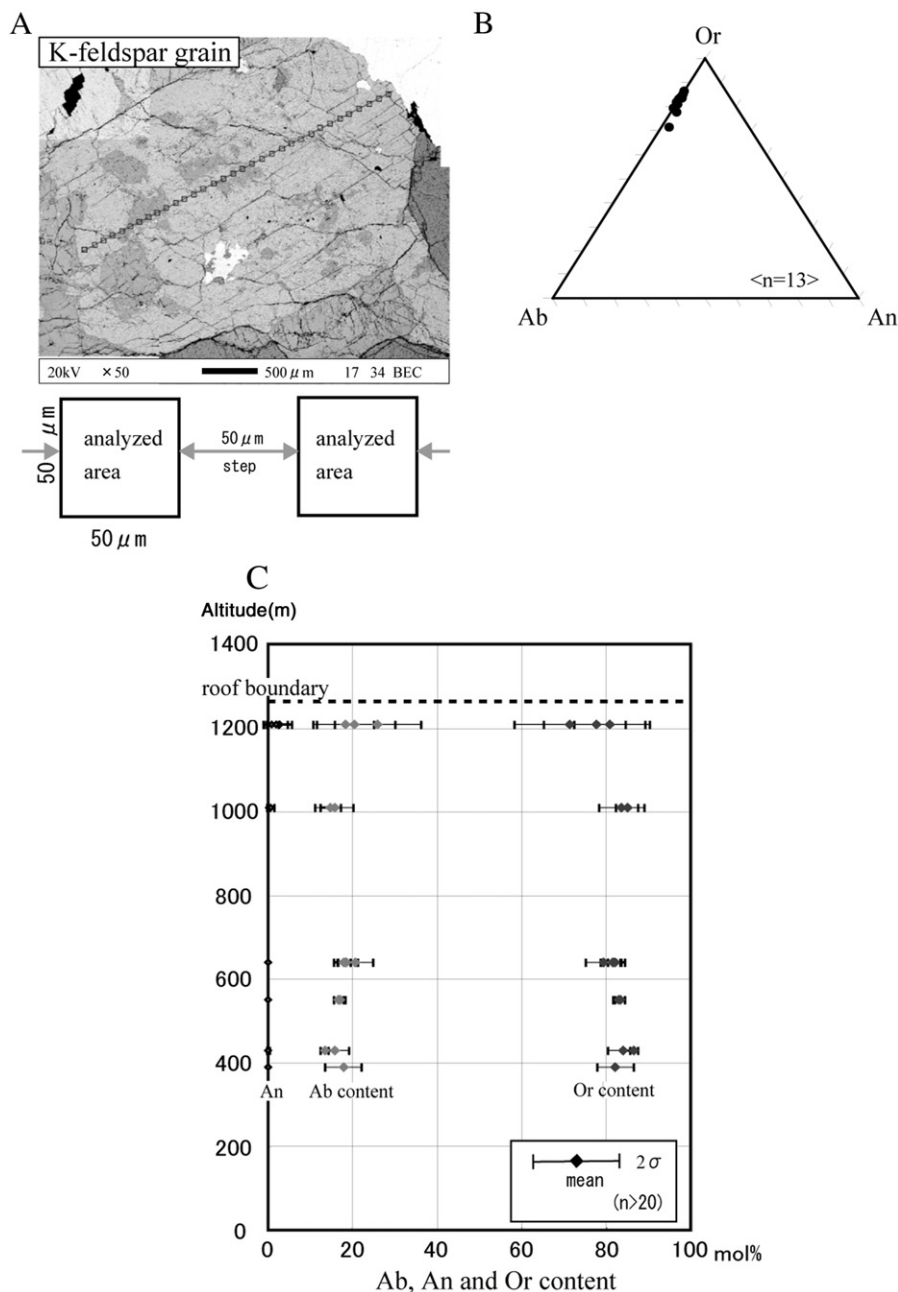


Fig. 13. Local bulk composition of K-feldspar grains. (A) Local bulk composition of K-feldspar determined as the mean composition in a 50 μm × 50 μm area taken at 50 μm intervals along an arbitrary line in perthitic K-feldspar grains (No. 02052209B). (B) Plot of local bulk composition of K-feldspar grains on an Or–Ab–An diagram. (C) Plot of local bulk composition (Or, Ab, and An content) of K-feldspar grains with altitude.

3.4.1. Volume fraction of albite-rich phase in perthitic K-feldspar

The volume fraction of albite lamellae in perthitic K-feldspar was estimated from the areal fraction of the

albite-rich phase in perthite grains assuming equality between the volume and areal fractions. After eliminating other minerals outside the K-feldspar grain from the BSE image, the albite-rich phase and host orthoclase-

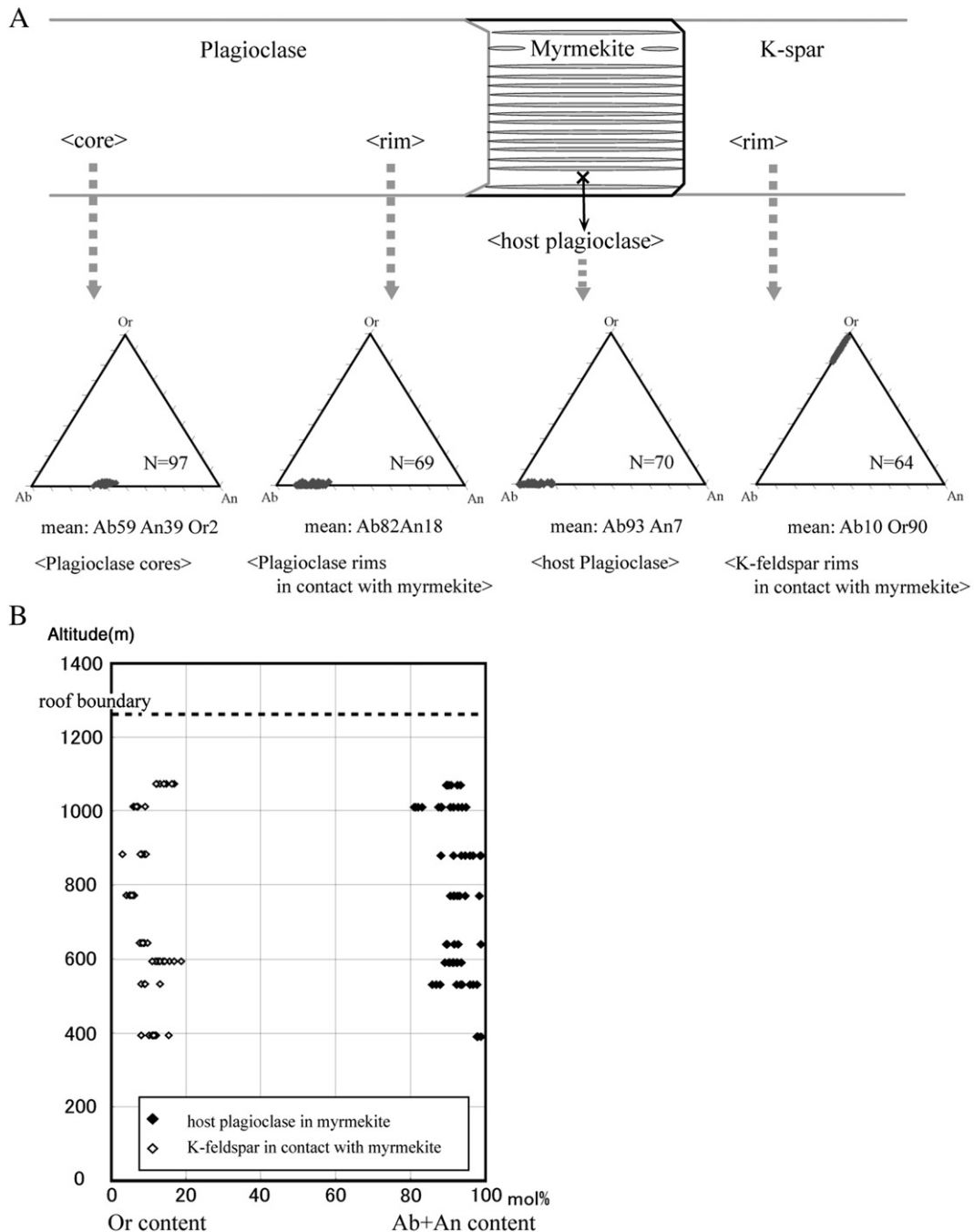


Fig. 14. Composition of plagioclase in myrmekite and neighboring minerals. (A) Or–Ab–An compositional plot of core and rim plagioclase in contact with myrmekite, host plagioclase in myrmekite, and K-feldspar rim in contact with myrmekite. (B) Compositional plot of host plagioclase in myrmekite and K-feldspar rim in contact with myrmekite against altitude (Or and Ab+An content).

rich phase inside the K-feldspar grain were converted to a binary image using image processing software (Scion image; Fig. 10(A)). The mean volume fraction of albite lamellae in perthitic K-feldspar was thus found to be an almost constant value of 0.30 (Fig. 10(B)).

3.4.2. Composition of albite lamellae and orthoclase-rich phase in microperthite

The composition of albite lamellae and the orthoclase-rich mother phase in microperthite was measured at points equally distant from the interface between the

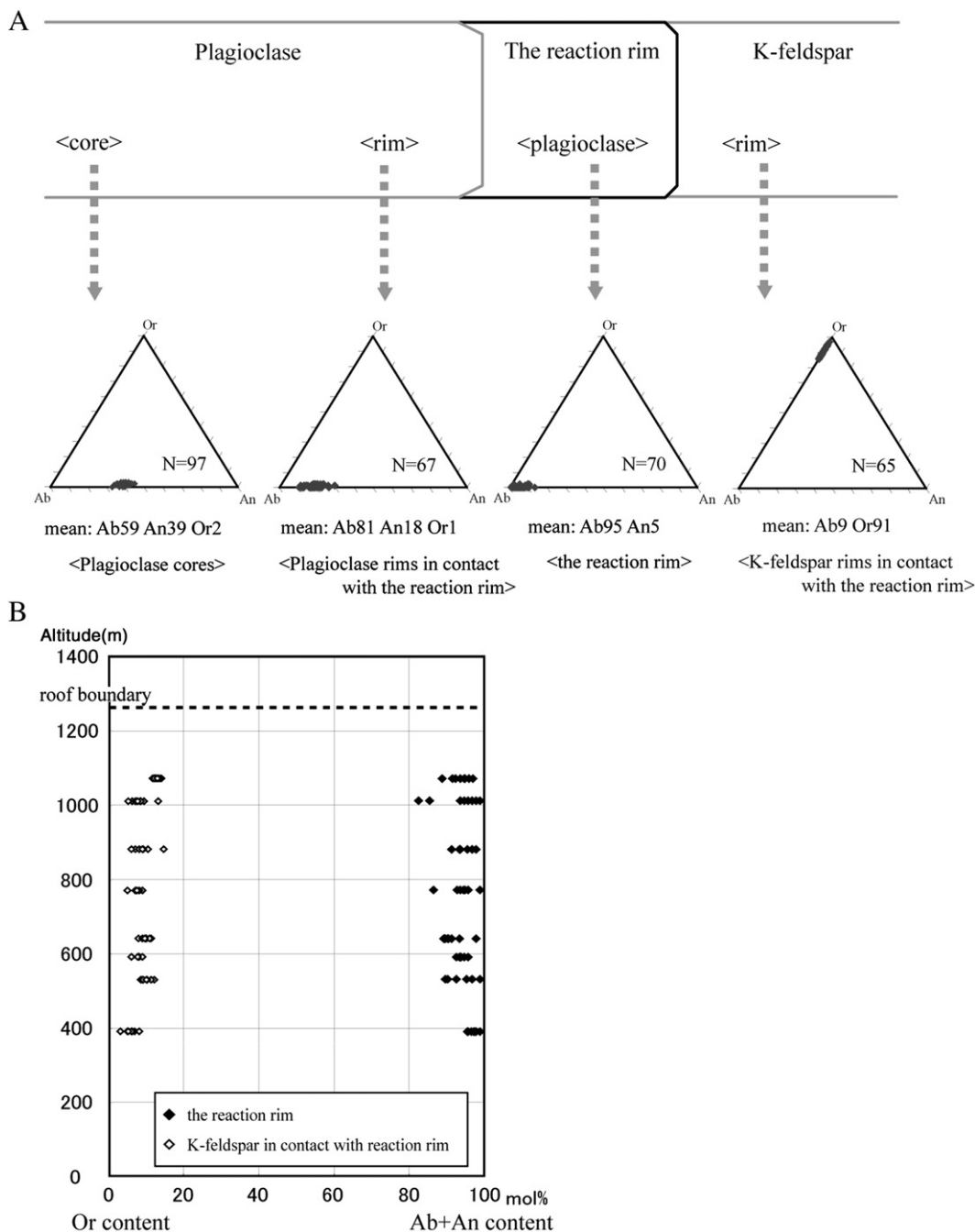


Fig. 15. Composition of reaction rims and neighboring minerals (Or, Ab, and An content). (A) Or–Ab–An compositional plot of core and rim plagioclase in contact with reaction rims, reaction rims, and K-feldspar rims in contact with reaction rims. (B) Compositional plot of reaction rim and K-feldspar rim in contact with reaction rim (Or and Ab+An content) against altitude.

two phases (Fig. 11(A)). The orthoclase-rich phase in microperthite is homogeneous, having a mean composition of $\text{Or}_{85}\text{Ab}_{15}$ ($\text{K}_{0.85}\text{Na}_{0.15}\text{AlSi}_3\text{O}_8$), whereas the composition of albite lamellae ranges from Ab_{61} to Ab_{85} with a mean composition of $\text{Ab}_{75}\text{Or}_{24}\text{An}_1$ ($\text{Na}_{0.75}\text{K}_{0.24}\text{Ca}_{0.01}\text{Al}_{1.01}\text{Si}_{2.99}\text{O}_8$; Fig. 11(B)). Microperthite lamellae thus contain a small amount of anorthite, which may be incorporated by deuteric alteration subsequent to exsolution. Fig. 11(C) shows composition plots for albite lamellae the orthoclase-rich phase (Or and Ab+An content) in microperthite with respect to altitude, revealing that the compositions are relatively uniform throughout the body.

3.4.3. Composition of albite lamellae and orthoclase-rich phase in patchperthite

Chemical analyses of albite lamellae and the orthoclase-rich phase in patchperthite were made at points equally distant from the interface between the two phases (Fig. 12(A)). The orthoclase-rich phase in patchperthite is homogeneous, exhibiting a mean composition of $\text{Or}_{90}\text{Ab}_{10}$ ($\text{K}_{0.90}\text{Na}_{0.10}\text{AlSi}_3\text{O}_8$), whereas the composition of albite lamellae varies from Ab_{78} to Ab_{99} with a mean composition of $\text{Ab}_{89}\text{Or}_5\text{An}_6$ ($\text{Na}_{0.89}\text{K}_{0.05}\text{Ca}_{0.06}\text{Al}_{1.06}\text{Si}_{2.94}\text{O}_8$; Fig. 12(B)). The patchperthite lamellae contain somewhat more anorthite than the microperthite lamellae, which may be due to alteration by deuteric fluids. Fig. 12(C) shows compositional plots for albite lamellae and the orthoclase-rich phase (Or and Ab+An content) in patchperthite, again revealing that the compositions are relatively uniform with respect to depth in the granite body.

3.4.4. Local bulk composition of K-feldspar grains

The local bulk composition of K-feldspar was measured as the mean composition in a $50\text{ }\mu\text{m} \times 50\text{ }\mu\text{m}$ area, with measurements taken at $50\text{ }\mu\text{m}$ intervals along an arbitrary line in each perthitic K-feldspar grain (Fig. 13(A)) following the technique of Hashimoto et al. (2005b). The local bulk composition of K-feldspar grains is homogeneous throughout the granite body, having a mean composition of $\text{Or}_{81}\text{Ab}_{18}\text{An}_1$ ($\text{K}_{0.81}\text{Na}_{0.18}\text{Ca}_{0.01}\text{Al}_{1.01}\text{Si}_{2.99}\text{O}_8$; Fig. 13(B)). The compositional plot confirms that the composition is almost constant with respect to altitude (Fig. 13(C)).

3.4.5. Composition of plagioclase in myrmekite and neighboring minerals

Fig. 14(A) shows a compositional plot (Or–Ab–An) of core and rim plagioclase in contact with myrmekite, host plagioclase in myrmekite, and the rim on K-feldspar in contact with myrmekite. The ternary diagram indicates

near-uniform compositions for each of the phases. Myrmekitic host plagioclase compositions (Or and Ab+An content; Fig. 14(b-1)) and K-feldspar rim compositions in contact with myrmekite (Fig. 14(b-2)) are also constant with respect to altitude.

3.4.6. Composition of reaction rims and neighboring minerals

Fig. 15(A) shows that the core and rim compositions of plagioclase in contact with the reaction rim, the compositions of the reaction rim, and the rim compositions of K-feldspar in contact with the reaction rim are nearly constant. The reaction rim composition (Or and Ab+An content; Fig. 15(b-1)) and K-feldspar rim composition in contact with the reaction rim (Fig. 15(b-2)) are also constant with respect to altitude.

Above results indicate that the chemical compositions of feldspars including the sub-solidus reaction textures do not show any systematic variations against the altitude, therefore we conclude that the bulk compositions of feldspars have almost constant compositions against the altitude.

4. Discussion

All six of the textural indicators display a systematic increasing trend with depth in the granite body, without obvious horizontal variations. In contrast, the compositions of each of the components are relatively uniform with depth. This result suggests that the Okueyama granite was effectively cooled not from the wall but from the roof, as the extent of sub-solidus reactions reflects the duration of annealing during which exsolution coarsening and deuteric coarsening can occur.

4.1. Temperature conditions of exsolution coarsening reactions

Estimation of the temperature regime in the exsolution coarsening stage is based on a combination of the hornblende–plagioclase thermometer and the ternary feldspar thermometer. The hornblende–plagioclase thermometer is useful here because the mineral pair occurs commonly in both the hornblende–biotite granite and hornblende–biotite granodiorite of the Okueyama granite. Hornblende in contact with plagioclase is essentially homogeneous except at the rims, which are up to $100\text{ }\mu\text{m}$ in thickness and exhibit increasing SiO_2 , MgO , and CaO compositions and decreasing TiO_2 , Al_2O_3 , FeO , Na_2O and K_2O compositions (Fig. 16). The core compositions of hornblende vary over a small range from ferroedenite to edenite without any dependence on

altitude (Fig. 17(A)). Recalculation of the ferrous and ferric irons in amphibole follows the methods of Holland and Blundy (1994; appendix B) and Spear and Kimball (1983). Rim compositions exhibit a much wider distribution in the amphibole classification diagram (Leake et al., 1997), covering ferroedenite, edenite, actinolite, ferro-actinolite, ferro-hornblende, and magnesio-hornblende (Fig. 17(B)). Plagioclase in contact with hornblende is also homogeneous, displaying a mean composition of An₃₈, although the rims (50–100 μ m) have a composition of An₂₂ (Fig. 17(C)). Neither hornblende nor plagioclase display compositional changes at the rims in contact with quartz. The

compositional changes at the rims of hornblende and plagioclase in contact with each other therefore suggest sub-solidus re-equilibration between the two phases by volume diffusion. The homogeneous core compositions thus reflect the compositions at crystallization, and the temperature calculated using the core composition will give both the crystallization temperature and the onset temperature of volume diffusion.

Among the three hornblende–plagioclase thermometers available (Blundy and Holland, 1990, 1994), only the thermometer of Blundy and Holland (1990) is applicable to the Okueyama granite, since the other two give erroneous results for iron-rich samples (Holland

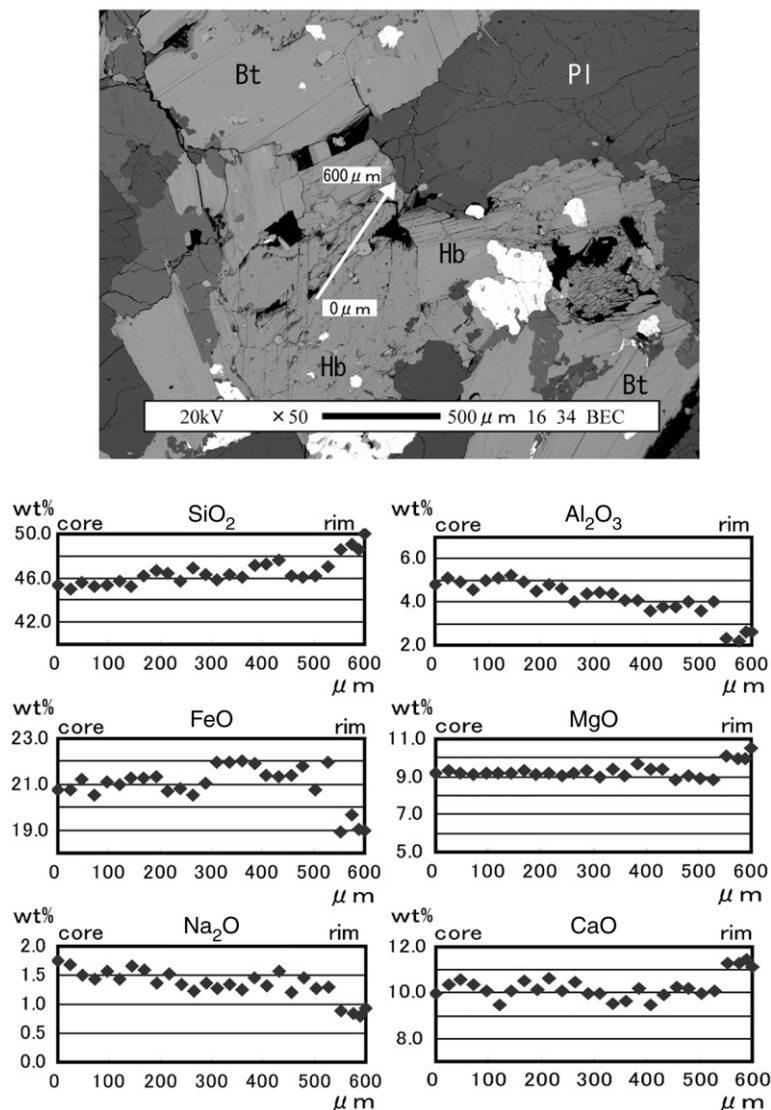


Fig. 16. BSE image (upper) of hornblende (arrow indicates line scan from core to rim) and corresponding compositional profile (lower) for sample No. 02041603 (altitude, 550 m).

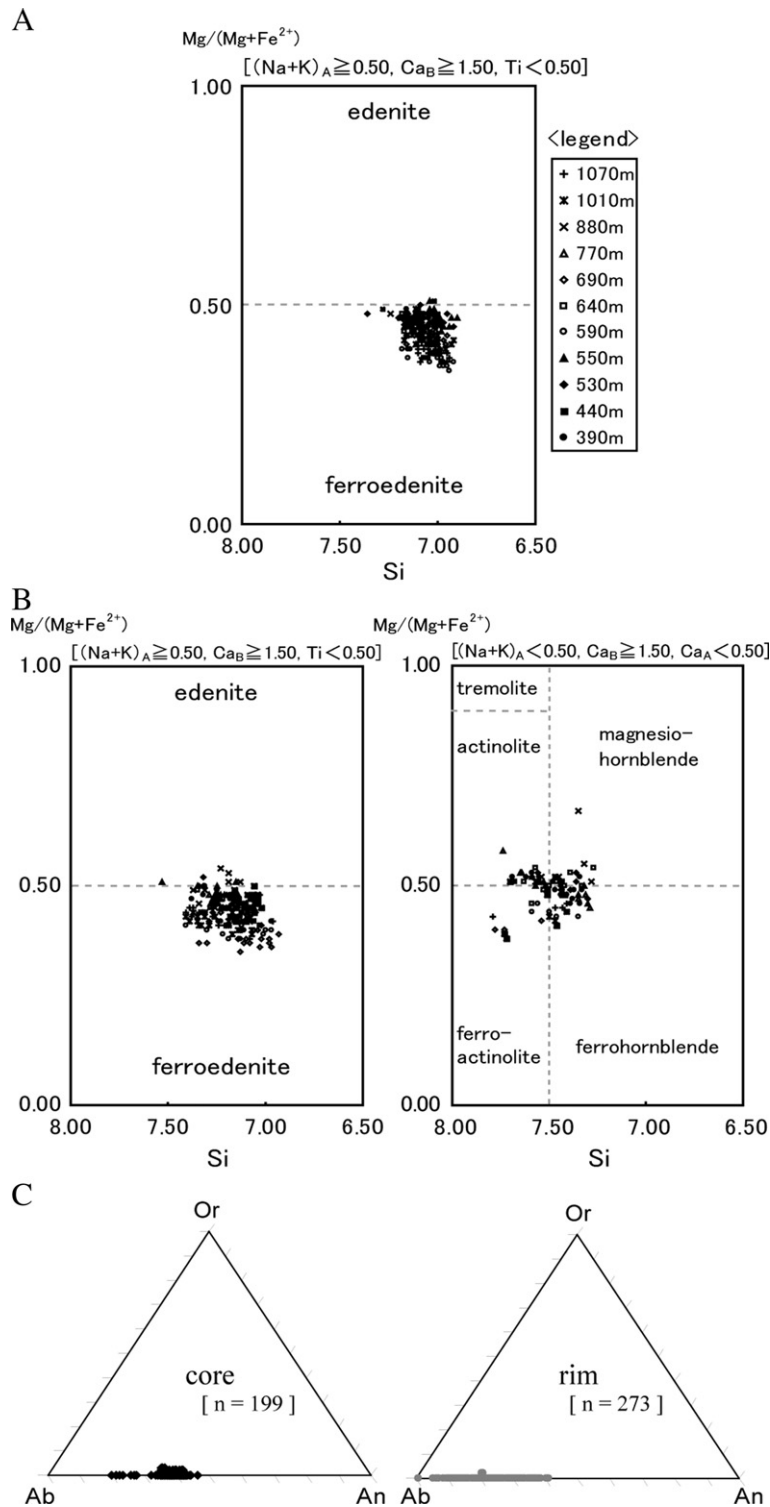


Fig. 17. Chemical composition of amphibole and plagioclase. (A,B) Plot of core (A) and rim (B) compositions of amphibole on classification diagram of Leake et al. (1997). (C) Or–Ab–An compositional plot of core and rim compositions of plagioclase in contact with hornblende.

Table 1
Chemical compositions of hornblende and plagioclase at cores from OKG

Sample no.	02041830	02041824	02041824	02091607	02091607	02091417	02041603	02051702	02041912
	A01	D03	D04	A05	A06	D04	A02	D11	B04
	Homblende	Homblende	Homblende	Homblende	Homblende	Homblende	Homblende	Homblende	Homblende
	Core	Core	Core	Core	Core	Core	Core	Core	Core
Altitude	1010 m	880 m	880 m	690 m	690 m	640 m	550 m	530 m	440 m
(wt.%)									
SiO ₂	45.63	45.04	45.60	45.14	44.57	45.12	45.01	47.62	45.35
TiO ₂	0.63	0.94	1.26	0.83	0.92	0.99	0.88	0.37	1.27
Al ₂ O ₃	4.55	4.96	5.98	5.87	5.67	5.28	5.14	3.99	6.06
FeO	21.62	20.57	20.95	22.29	21.86	22.35	20.74	20.94	20.47
MnO	0.77	0.84	0.50	1.30	1.11	0.82	0.71	0.68	0.47
MgO	8.73	9.13	9.43	8.05	7.95	8.52	9.33	9.85	9.76
CaO	11.26	10.00	10.39	10.45	10.49	10.33	10.35	10.86	10.06
Na ₂ O	1.48	1.62	1.60	1.67	1.66	1.44	1.69	1.09	1.49
K ₂ O	1.4	1.51	1.77	1.59	1.75	1.87	1.22	0.81	1.71
Total	96.07	94.61	97.48	97.19	95.98	96.72	95.07	96.21	96.64
(atom)									
Si	7.17	7.14	7.02	7.04	7.04	7.07	7.10	7.36	7.00
Ti	0.07	0.11	0.15	0.10	0.11	0.12	0.10	0.04	0.15
Al ^{IV}	0.83	0.86	0.98	0.96	0.96	0.93	0.90	0.64	1.00
Al ^{VI}	0.01	0.07	0.10	0.12	0.10	0.04	0.06	0.09	0.10
Fe ³⁺	0.03	0.20	0.28	0.34	0.28	0.11	0.17	0.26	0.28
Fe ²⁺	2.81	2.52	2.40	2.55	2.59	2.81	2.56	2.44	2.38
Mn	0.10	0.11	0.06	0.17	0.15	0.11	0.09	0.09	0.06
Mg	2.04	2.16	2.16	1.87	1.87	1.99	2.19	2.27	2.25
Ca	1.90	1.70	1.71	1.75	1.78	1.73	1.75	1.80	1.66
Na	0.45	0.50	0.48	0.50	0.51	0.44	0.52	0.33	0.45
K	0.28	0.31	0.35	0.32	0.35	0.37	0.24	0.16	0.34
Total	15.69	15.69	15.71	15.74	15.76	15.73	15.69	15.49	15.69
Mg#	0.42	0.46	0.47	0.42	0.42	0.41	0.46	0.48	0.48
Sample no.	A01	D03	D04	A05	A06	D04	A02	D11	B04
	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase
	Core	Core	Core	Core	Core	Core	Core	Core	Core
(wt.%)									
SiO ₂	58.01	57.13	57.50	55.65	56.94	57.91	57.64	56.83	58.92
Al ₂ O ₃	27.88	26.22	25.93	28.06	28.13	27.29	27.76	28.47	25.96
CaO	7.67	8.26	7.88	7.67	7.96	7.02	7.74	7.98	7.37
Na ₂ O	7.08	7.16	7.38	7.00	6.77	7.33	7.09	6.78	7.33
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.27	0.00
Total	100.64	98.77	98.69	98.38	99.80	99.55	100.20	100.33	99.58
(atom)									
Si	2.57	2.59	2.61	2.53	2.55	2.59	2.57	2.54	2.64
Al	1.46	1.40	1.39	1.50	1.48	1.44	1.46	1.50	1.37
Ca	0.37	0.40	0.38	0.37	0.38	0.34	0.37	0.38	0.35
Na	0.61	0.63	0.65	0.62	0.59	0.64	0.61	0.59	0.64
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.00
Total	5.00	5.02	5.02	5.02	5.00	5.01	5.01	5.03	5.00
Ab	0.622	0.612	0.631	0.626	0.608	0.653	0.622	0.596	0.646
An	0.378	0.388	0.369	0.374	0.392	0.347	0.378	0.384	0.354
Or	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Mg# = Mg / (Mg + Fe ²⁺)									

Table 2

Chemical compositions of hornblende and plagioclase at rims from OKG

Sample no.	02041830	2041824	02041824	02091417	02091417	02051702	02051702	02041912	02041912
	A01	C01	E03	A05	D06	A03	B01	B02	C09
	Hornblende	Hornblende	Hornblende	Hornblende	Hornblende	Hornblende	Hornblende	Hornblende	Hornblende
	Rim	Rim	Rim	Rim	Rim	Rim	Rim	Rim	Rim
Altitude	1010 m	880 m	880 m	640 m	640 m	530 m	530 m	440 m	440 m
(wt.%)									
SiO ₂	48.18	49.99	46.82	50.57	47.18	47.82	45.35	45.73	45.95
TiO ₂	0.30	0.37	0.51	0.09	0.82	0.84	1.11	0.95	1.06
Al ₂ O ₃	3.22	3.01	4.11	2.47	5.01	4.13	4.90	5.18	5.92
FeO	22.33	19.17	19.78	19.53	21.13	18.72	21.55	20.77	20.04
MnO	1.00	0.45	0.75	0.56	0.53	0.56	0.71	0.76	0.83
MgO	8.48	9.97	9.70	10.74	9.18	10.41	9.62	8.94	9.64
CaO	11.56	11.52	11.25	12.12	10.43	10.84	10.26	10.72	10.55
Na ₂ O	0.96	0.97	1.04	0.70	1.58	1.27	1.07	1.34	1.65
K ₂ O	0.63	0.52	1.53	0.40	1.53	1.06	1.89	1.23	1.48
Total	96.66	95.97	95.49	97.18	97.39	95.65	96.46	95.62	97.12
(atom)									
Si	7.47	7.52	7.31	7.63	7.24	7.36	7.09	7.16	7.06
Ti	0.03	0.04	0.06	0.01	0.09	0.10	0.13	0.11	0.12
Al ^{IV}	0.53	0.48	0.69	0.37	0.76	0.64	0.91	0.84	0.94
Al ^{VI}	0.06	0.08	0.07	0.07	0.15	0.11	0.00	0.12	0.13
Fe ³⁺	0.17	0.21	0.20	0.20	0.39	0.31	0.00	0.34	0.37
Fe ²⁺	2.71	2.29	2.38	2.26	2.30	2.08	2.82	2.36	2.19
Mn	0.13	0.06	0.10	0.07	0.07	0.07	0.09	0.10	0.11
Mg	1.96	2.33	2.26	2.42	2.10	2.39	2.24	2.08	2.21
Ca	1.92	1.94	1.88	1.96	1.71	1.79	1.72	1.80	1.74
Na	0.29	0.29	0.31	0.20	0.47	0.38	0.32	0.41	0.49
K	0.12	0.10	0.30	0.08	0.30	0.21	0.38	0.25	0.29
Total	15.40	15.35	15.57	15.28	15.60	15.46	15.69	15.59	15.67
Mg#	0.42	0.50	0.48	0.51	0.47	0.53	0.44	0.47	0.50
Sample no.	A01	C01	E03	A05	D06	A03	B01	B02	C09
	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase	Plagioclase
	Rim	Rim	Rim	Rim	Rim	Rim	Rim	Rim	Rim
(wt.%)									
SiO ₂	61.37	62.70	62.45	63.68	64.28	63.45	61.87	62.07	61.33
Al ₂ O ₃	23.79	24.33	23.38	23.08	23.89	23.28	23.26	23.03	23.83
CaO	5.40	4.76	4.92	3.84	4.20	4.14	4.71	4.36	6.24
Na ₂ O	8.25	8.95	8.99	9.46	9.33	9.54	9.03	8.80	7.66
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	98.81	100.74	99.74	100.06	101.70	100.41	98.87	98.30	99.06
(atom)									
Si	2.75	2.75	2.77	2.81	2.79	2.79	2.77	2.79	2.71
Al	1.26	1.26	1.22	1.20	1.22	1.21	1.23	1.22	1.29
Ca	0.26	0.22	0.23	0.18	0.20	0.20	0.23	0.21	0.31
Na	0.72	0.76	0.77	0.81	0.78	0.81	0.78	0.77	0.68
K	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	4.99	4.99	4.99	5.00	4.99	5.01	5.01	4.99	4.99
Ab	0.735	0.776	0.770	0.818	0.796	0.802	0.772	0.786	0.687
An	0.265	0.224	0.230	0.182	0.204	0.198	0.228	0.214	0.313
Or	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Mg# = Mg / (Mg + Fe ²⁺)									

Table 3

The mean and maximal crystallization pressure of OKG evaluated by Al-in-hornblende barometer (Schmidt, 1992)

Sample no.	Altitude (m)	Number ^a	Pressure (MPa)	
			Mean	Maximum
02052405	1070	17	170	213
02041830	1010	19	197	256
02041824	880	19	160	213
02102302	770	14	144	164
02091607	690	15	187	223
02091417	640	20	165	218
02051706	590	20	198	237
02041603	550	21	223	294
02051702	530	17	173	251
02041912	440	13	180	227
02071415	390	24	183	227

^a The number of analyses.

and Blundy, 1994), as in the present case. The applicability of Blundy and Holland's (1990) thermometer is restricted to amphiboles with $\text{Si} < 7.8$ and to plagioclase with $\text{An} < 92$, both of which are satisfied in the Okueyama samples (Tables 1 and 2). As the thermometer requires knowledge of pressure, one of the aluminum-in-hornblende barometers (Hammerstrom and Zen, 1986; Holister et al., 1987; Johnson and Rutherford, 1988; Blundy and Holland, 1990; Schmidt, 1992; Anderson and Smith, 1995) is employed here. Of the several barometers available, that of Schmidt (1992) is adopted as being suitable for a water-saturated condition, which is compatible with the Okueyama granite (Takahashi, 1986).

Table 3 shows the calculated mean and maximum pressures based on the core composition of hornblende. The differences between the mean and maximum values are approximately 50 MPa, which is within the error of the barometer. Therefore, the mean pressure of 180 MPa is adopted as the solidification pressure of the rock. This value is invariant with altitude.

Using this value of pressure, the solidification temperatures are calculated based on the core compositions of hornblende and plagioclase (Table 1). Fig. 18 shows a plot of the mean temperature with respect to altitude. The data reveal no systematic variation with altitude, indicating instead an almost constant value of ca. 710 °C. Also shown in the figure are the calculated temperatures based on the rim compositions of hornblende and plagioclase (Table 2), which give the terminal temperature for volume diffusion in minerals under the sub-solidus condition. Again, there is no systematic variation with altitude, and the temperature is a constant of 620 °C. The difference between the solidification temperature and the

terminal temperature for volume diffusion (ca. 90 °C) represents the temperature interval over which exsolution coarsening reactions took place.

The temperature conditions of the exsolution coarsening stage are evaluated independently by applying the ternary feldspar thermometer, which is based on the concept that the activities of albite, anorthite and orthoclase are identical in two coexisting feldspars (plagioclase and K-feldspar) in equilibrium under a given pressure and temperature condition. The thermometer is expressed as

$$a_{\text{Ab}}^{\text{Pl}} = a_{\text{Ab}}^{\text{Kfs}}$$

$$a_{\text{An}}^{\text{Pl}} = a_{\text{An}}^{\text{Kfs}}$$

$$a_{\text{Or}}^{\text{Pl}} = a_{\text{Or}}^{\text{Kfs}}$$

A pair of coexisting feldspars thus gives three temperature conditions (T_{Ab} , T_{An} , and T_{Or}) based on the activities of the constituent albite, anorthite, and orthoclase. Among the four ternary thermometers (Ghiorso,

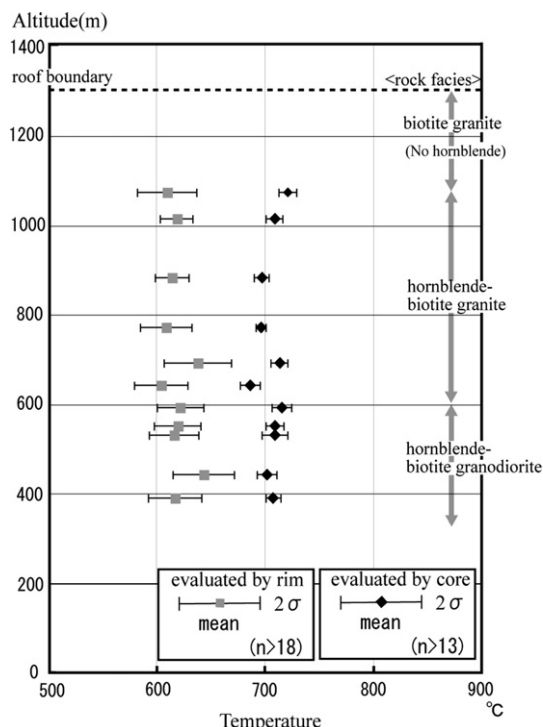


Fig. 18. Hornblende–plagioclase geothermometer (Blundy and Holland, 1990) giving crystallization temperature based on core compositions (diamonds) and re-equilibration temperature under sub-solidus exsolution condition based on rim compositions (squares).

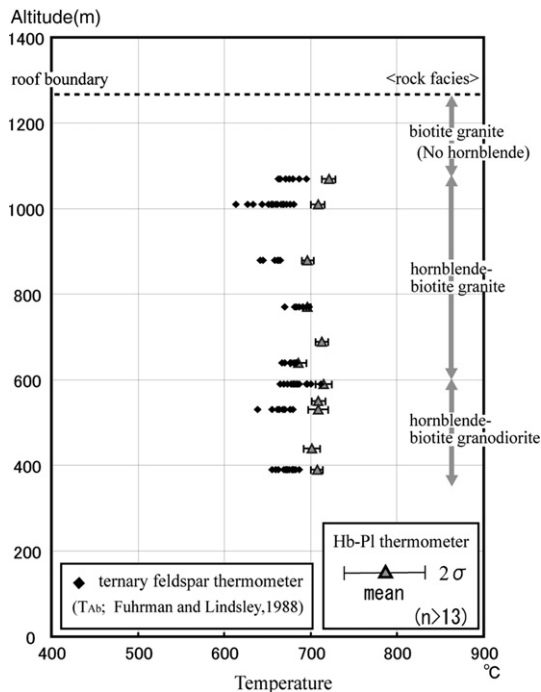


Fig. 19. Variation in crystallization temperature with altitude based on the ternary feldspar thermometer of Fuhrman and Lindsley (1988) using plagioclase core compositions and K-feldspar local bulk compositions (diamonds), and based on the hornblende–plagioclase thermometer of Blundy and Holland (1990) using core compositions (triangles).

1984; Green and Usdansky, 1986; Fuhrman and Lindsley, 1988; Elkins and Grove, 1990), the ternary feldspar thermometer of Fuhrman and Lindsley (1988) is applicable to the Okueyama granite, as this thermometer yields nearly concordant temperatures for plutonic igneous rocks (Fuhrman and Lindsley, 1988; Elkins and Grove, 1990). The binary thermometer of Brown and Parsons (1981) is also employed. Fuhrman and Lindsley (1988) mentioned the potential applicability of their thermometer to ‘close-to-equilibrium’ feldspar pairs that have undergone a sub-solidus reaction. The orthoclase temperature for metamorphic feldspar pairs and plutonic feldspar pairs tends to be lower than the temperatures of albite and anorthite, since T_{Or} is sensitive to re-equilibrium during the sub-solidus stage in slowly cooled rocks (Fuhrman and Lindsley, 1988). The temperature estimation for feldspar in the Okueyama granite will be predominantly dependent on the extent of albite equilibrium, since the orthoclase content in plagioclase and anorthite content in K-feldspar are small (Figs. 11–15).

If equilibrium between plagioclase and K-feldspar is established, the core compositions of both minerals can

be used for estimation of the crystallization temperature. However, as the original core composition of K-feldspar have been modified by exsolution and deuteric alteration, the local bulk composition of K-feldspar ($Or_{81}Ab_{18}An_1$; Fig. 13(B)), which implies the original (pristine) composition, is used for estimation of the crystallization temperature. Fig. 19 shows a comparison of the crystallization temperature estimated by the ternary feldspar thermometer with that by the hornblende–plagioclase thermometer based on core compositions. While the former approach gives a range of 710–610 °C with altitude, the temperatures are up to 80 °C lower than those given by the latter thermometer. The lower estimates may be due to poor estimates of the local bulk of K-feldspar, which cannot fully eliminate the effect of deuteric alteration. Thus, the highest temperature (711 °C) evaluated by the ternary feldspar thermometer is considered to be the most likely crystallization temperature, and is concordant with that estimated by the hornblende–plagioclase thermometer. Consequently, the two thermometers give a consistent result indicating an onset temperature of 710 °C for the exsolution reaction stage.

The terminal temperature of the ECS, corresponding to the re-equilibrium temperature of volume diffusion in

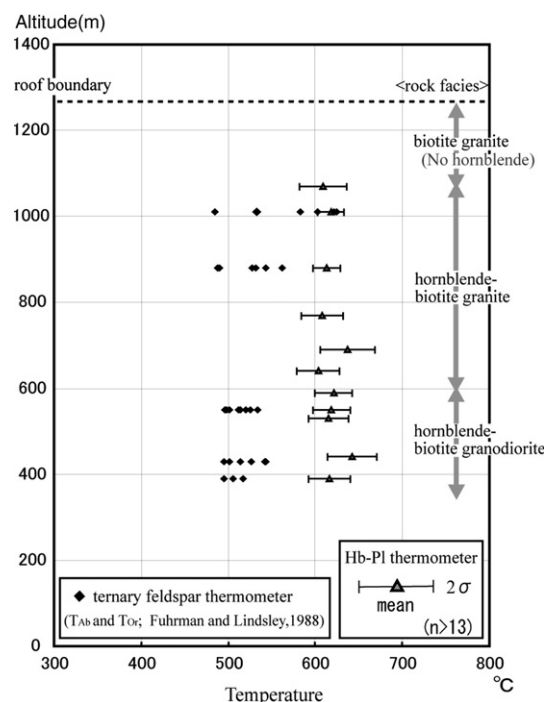


Fig. 20. Plot of exsolution coarsening terminal temperatures estimated by the ternary feldspar thermometer (diamonds) and the hornblende–plagioclase thermometer (triangles) with altitude.

minerals, is estimated by applying the ternary feldspar thermometer to the micropertite lamellae composition and the corresponding orthoclase-rich phase composition (see Fig. 11). Fig. 20 shows the terminal temperature of exsolution coarsening estimated by the ternary feldspar thermometer and the hornblende–plagioclase thermometer based on rim compositions. The exsolution terminal temperatures evaluated by the ternary feldspar thermometer vary from 620 °C to 500 °C with altitude, and are up to 100 °C lower than the temperatures given by the hornblende–plagioclase thermometer. The lower estimates by the ternary feldspar thermometer may also be due in this case to the influence of deuteritic alteration because micropertite lamellae can act as pathways for deuteritic fluids (Lee et al., 1995), as shown by the presence of etch-pits (Fig. 3(D)). Thus, the most likely terminal temperature of the ECS is the highest value (625 °C) given by the ternary feldspar thermometer, which is concordant with that determined by the hornblende–plagioclase thermometer. It is thus concluded that the terminal temperature of the ECS is ca. 620 °C.

4.2. Temperature conditions of deuteritic coarsening reactions

The temperature conditions of the DCS, under which patchperthite, myrmekite, and the reaction rims formed, are evaluated using the ternary feldspar thermometer of Fuhrman and Lindsley (1988). The estimated temperatures are below 500 °C for all three components. Under such low temperatures, an exact estimate of temperature is not possible due to ordering and kinetic effects. However, it can be noted that the DCS occurred at temperatures below 500 °C, representing substantially cooler conditions than those for the ECS.

4.3. Cooling periods corresponding to ECS and DCS

The period over which the Okueyama granite cooled from 710 °C to 620 °C can be considered to correspond to the period of sub-solidus exsolution coarsening. The cooling period is estimated here using a 1D numerical model of heat transfer. As it has been established in the present study that the Okueyama granite cooled from the

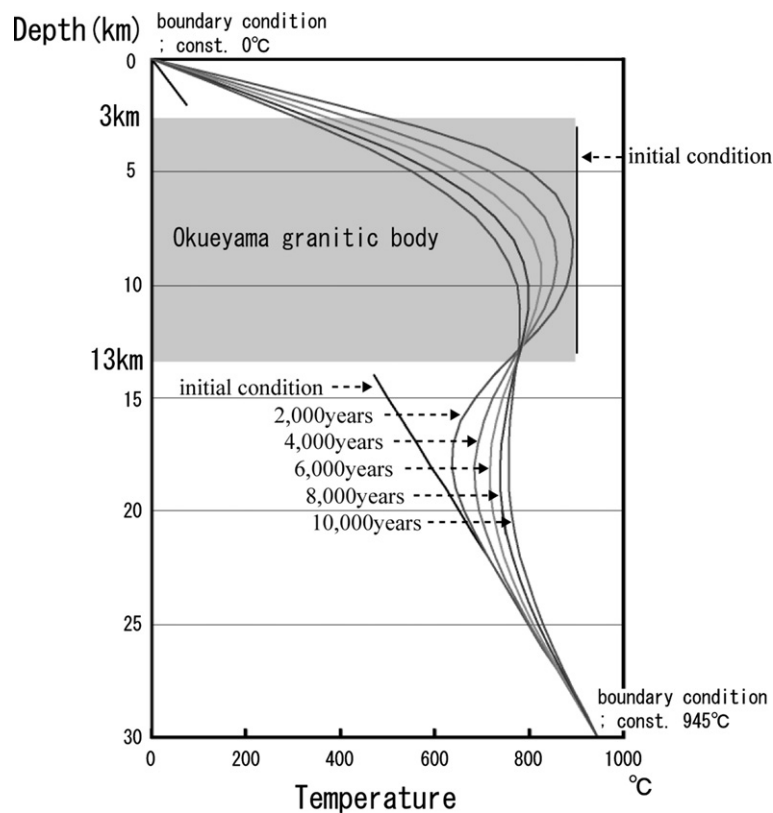


Fig. 21. Temperature–depth curves of thermal evolution for the 1D thermal model from the initial condition to 10,000 y with a time interval of 2000 y. The magma body is emplaced between 3 km (top) and 13 km (bottom).

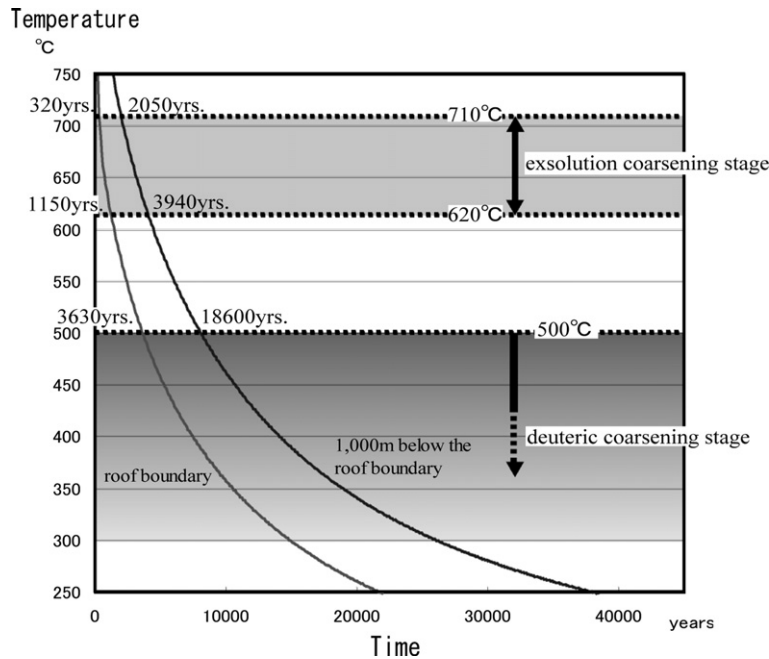


Fig. 22. Cooling period of exsolution coarsening stage (710–620 °C) at the roof boundary (3 km depth) and 1000 m below the roof (4 km depth) estimated by 1D thermal simulation with the magma body emplaced between 3 km and 13 km.

roof, the cooling process can be effectively described by a 1D heat transfer model using the following heat transfer equation (Turcotte and Schubert, 1982; Samarskii and Vabishchevich, 1995; Okudaira, 1996):

$$\rho_m C_m \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial z^2} + A$$

where ρ_m and C_m are the density (2750 kg/m³) and heat capacity (880 J/kg/K) of the rock, T is temperature, t is time, z is depth, K is thermal conductivity (2.8 W/m/K; Okudaira, 1996), and A is the heat production rate by radiometric decay in the crust. The density and heat capacity of the magma are assumed to be the same as those of the rock (Hanson and Barton, 1989; Rothstein and Hoisch, 1994; Okudaira, 1996).

The crust is assumed to be 30 km thick, with uniform temperatures at the surface ($T_s=0$ °C) and the base (945 °C). The initial temperature of the Okueyama magma is assumed to be 900 °C. The equation is solved using a finite difference scheme with $\Delta z=1$ km and $\Delta t=100$ y. The top of the magma body is emplaced at a subsurface depth of 3 km, and magma depths of 5–10 km are tested considering previous estimates of the depth of the Okueyama granite (>5 km; Takahashi, 1988).

Fig. 21 shows the results of 1D numerical simulations of heat transfer after intrusion of the Okueyama granite at depths of 3 km (top) and 13 km (bottom). The

figure shows the thermal evolution of the crust over 10 ka with a time interval of 2 ka. Thermal relaxation is thus relatively rapid in the present case. The cooling period corresponding to the ECS (710–620 °C) is evaluated at the roof and at the base of the exposure (1000 m below the roof). The period of the ECS is found to be relatively constant at 830 y regardless of the total depth of the magma body, whereas the period of the ECS at the base of the exposure varies considerably depending on the total body depth, from 1890 y for a 10 km-deep body (Fig. 22), to 1390 y for a 5 km-deep body (Table 4).

Table 4

The duration of cooling from the initial condition (900 °C) to 710 °C, and from that to 620 °C, and their difference (the cooling duration for temperature interval of 90 °C), at the roof boundary and at the 1000 m below the roof, estimated by 1-D heat transfer simulation with variable thickness of OKG from 5 to 10 km

The thickness in OKG (km)	At the roof boundary (year)			At 1000 m below the roof (year)		
	710 °C	620 °C	710 °C– 620 °C	710 °C	620 °C	710 °C– 620 °C
3–8	5	320	1140	820	1890	3280
3–9	6	320	1150	830	1990	3580
3–10	7	320	1150	830	2030	3760
3–11	8	320	1150	830	2040	3870
3–12	9	320	1150	830	2050	3920
3–13	10	320	1150	830	2050	3940

In the case of the DCS, the temperature at the roof reaches 500 °C after 3630 y assuming a 10 km-deep magma body, while the same temperature is reached after 8090 y at the bottom of the exposure (Fig. 22). However, as the temperature interval for DCS cannot be determined, the cooling period corresponding to the DCS is unknown.

The above results show that the duration of the ECS increases monotonically from the roof toward the bottom of the body. This finding is consistent with the results for the Klokken gabbro–syenite reported by Brown and Parsons (1984), which indicate a reaction period of several years at the top of the body and several hundreds to one thousand years at a depth of 600 m.

5. Conclusion

The Okueyama granite, a zoned granitic pluton, exhibits sub-solidus reaction textures formed by exsolution and deuteric coarsening. The extent of these textures increases systematically from the roof of the body to the base of the exposure. The extents of the two stages of sub-solidus reactions therefore represent good indicators of the cooling periods corresponding to this reaction activity, providing a means of constraining the model of magmatic cooling for this body.

The temperatures of initiation and termination of exsolution coarsening reactions were estimated using a combination of the hornblende–plagioclase thermometer of Blundy and Holland (1990) and the ternary feldspar thermometer of Fuhrman and Lindsley (1988) to be 710 °C and 620 °, respectively. The thermal conditions of deuteric coarsening reactions were evaluated using the ternary feldspar thermometer of Fuhrman and Lindsley (1988), suggesting temperatures below 500 °C. Numerical simulation of heat transfer using a 1D model indicates that cooling from 710 °C to 620 °C, corresponding to the period of exsolution reactions, occurred over 830 y at the roof and 1390–1890 y at the base of the exposure (1000 m below the roof), depending on the total depth of the original magma body. That is, the period of cooling corresponding to the exsolution coarsening increased with depth. This case study employing both petrographical and numerical methods provides useful information on the thermal evolution of the crust in the outer zone of the Japanese Islands, where granitic intrusions occur sporadically.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at [doi:10.1016/j.lithos.2007.01.005](https://doi.org/10.1016/j.lithos.2007.01.005).

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