Granulite facies paragneisses from the middle segment of the Mogok metamorphic belt, central Myanmar

Ye Kyaw Thu*, Masaki ENAMI**, Takenori KATO** and Motohiro TSUBOI***

*Department of Earth and Planetary Sciences, Nagoya University, Nagoya 464-8601, Japan
**Institute for Space–Earth Environmental Research, Nagoya University, Nagoya 464-8601, Japan
***Department of Applied Chemistry for Environment, School of Science and Technology, Kwansei Gakuin University, Sanda, Hyogo 669-1337, Japan

The Mogok metamorphic belt in central Myanmar is composed mainly of high-temperature paragneisses, marbles, calc-silicate rocks, and granitoids. The garnet-biotite-plagioclase-sillimanite-quartz and garnet-cordierite-sillimanite-biotite-quartz assemblages and their partial systems suggest pressure-temperature (P-T) conditions of 0.60–0.79 GPa/800–860 °C and 0.65 GPa/820 °C, respectively, for the peak metamorphic stage, and 0.40 GPa/620 °C for the exhumation stage. Ti-in-biotite and Zr-in-rutile geothermometers also indicate metamorphic equilibrium under upper amphibolite- and granulite facies conditions. Comparison of these estimates with previously described P-T conditions suggests that (1) the metamorphic conditions of the Mogok metamorphic belt vary from the lower amphibolite- to granulite facies, (2) metamorphic grade seems to increase from east to west perpendicular to the north-trending extensional direction of the Mogok belt, (3) granulite facies rocks are widespread in the middle segment of the Mogok belt, and (4) the granulite facies rocks were locally re-equilibrated at lower amphibolite facies conditions during the exhumation.

Keywords: P-T conditions, Paragneiss, Mogok metamorphic belt, Myanmar

INTRODUCTION

The Mesozoic to Cenozoic tectonic evolution of Southeast Asia is characterized by the closure of the Tethyan oceans and subsequent collision of the Indian microcontinent with Eurasia (Metcalfe, 1994, 2013). Progressive convergence between these two continental plates resulted in the development of Cenozoic metamorphism and related magmatism in central Myanmar that consists of Shan–Thai (Sibumasu) Block and West Burma Block in the east and west, respectively (Bertrand et al., 1999; Barley et al., 2003; Searle et al., 2007). The Cenozoic Mogok metamorphic belt is located at the western margin of the Shan–Thai Block south of the eastern Himalayan syntaxis; it forms a prominent tectonic sliver that is important for understanding the continental evolution of Southeast Asia (Fig. 1a). Systematic investigations of tectonometamorphic conditions of the Mogok metamorphic belt and related magmatism are therefore essential to the investigation of tectonic evolution of this region.

The Mogok metamorphic belt extends about 1500 km from south to north, but recent studies have focused entirely on the middle segment of the belt in the Mogok–Mandalay region (Fig. 1b). Geochronological studies have indicated, based on U-Th-Pb and Ar-Ar dating, that an assemblage of the Mogok high-grade metamorphic rocks formed during the Paleogene to early Neogene in association with the India–Eurasia continental collision (Bertrand et al., 1999; Bertrand et al., 2001; Maw Maw Win et al., 2016). Peak metamorphic conditions appear to vary between different parts of the elongated belt, such as the amphibolite facies in the Kyanigan and Kyauske areas (e.g., Searle et al., 2007) and the granulite facies in the northern Mogok region (Yonemura et al., 2013) and the Sagaing ridge (Maw Maw Win et al., 2016). Ye Kyaw Thu et al. (2016) reported a paragneiss with a spinel + quartz assemblage coexisting with Ti-rich biotite (up to 6.9 wt% TiO_2) that formed under granulite facies conditions. These results imply that high-grade metamorphic rocks occur extensively in the western Shan–Thai Block. However, mineralogical and petrological charac-
teristics of the Mogok metamorphic rocks, including pressure ($P$)–temperature ($T$) variations and $P$–$T$ trajectory, which are important for understanding the evolution of continental crust in Southeast Asia, have been poorly constrained.

We applied Ti–in–biotite and Zr–in–rutile geothermometers together with conventional geothermobarometers for the granulite facies rocks that occur commonly in the middle segment of the Mogok metamorphic belt. Our results yield basic information necessary for understanding the tectonometamorphic evolution of Southeast Asia.

GEOLOGICAL SETTING

Myanmar is geologically divided into two provinces, the Eastern and Western Provinces, separated by the N–S trending Sagaing dextral strike–slip fault (Fig. 1b). The Western Province, or the Burma microplate, includes the Indo–Burma ranges and Central basins (Fig. 1b). The Indo–Burma ranges, where the Burma microplate overrides Indian oceanic crust, consists of Late Cretaceous–Paleogene marine sedimentary rocks overlying flysch-type sediment, and is coupled to the east with a discontinuous linear belt of ophiolitic rocks that mark the final closure of the Neo–Tethys between India and Southeast Asia (Acharyya, 2007; Ghose et al., 2014; Gardiner et al., 2015). The Central basins are covered by a series of Cretaceous to Pleistocene sediments and Quaternary volcanic rocks (Mitchell, 1993). The boundary between the Eastern and Western Provinces is marked by the 1200-km–long Sagaing dextral fault that splays off in the north to form three metamorphic ranges: the jade mine uplift, the Katha–Gangaw range, and the Tagaung–Myitkyina belt (Fig. 1b). The jade mine uplift consists of jadeite–bearing rocks with serpentinized peridotite (Shi et al., 2012), which have been metamorphosed at eclogite facies conditions of $P > 1.4$ GPa and $T \sim 550–600$ °C (Goffé et al., 2002). The Katha–Gangaw range and Tagaung–Myitkyina belt, however, are composed mainly of low– to medium–pressure/high–temperature metamorphic rocks, which are most likely a continuation of the Mogok metamorphic belt (Mitchell et al., 2007). Enami et al. (2012) reported an eclogite occurring at the boundary zone (Kumon range) between the jade mine uplift and Katha–Gangaw range; this occurrence likely belongs to the jade mine uplift.

The Eastern Province is a part of the Shan–Thai Block that collided with the Indochina Block in the Early or Late Triassic (Mitchell, 1977; Metcalfe, 1988; Metcalfe, 2000). It comprises the Shan Plateau, the Mergui Group, and the Mogok metamorphic belt (Fig. 1b). Most of the Shan–Thai Block is characterized by an Upper Cambrian to Triassic succession (Boucot, 2002) overlying metamorphic rocks of the Chaung Magyi Group. The Mergui Group, also called the Slate Belt (Mitchell, 1992), consists of distinctive Late Paleozoic glacial marine peb-
bly mudstones that extend to Thailand, the Malay Peninsula and Sumatra (Barber and Crow, 2009; Ridd and Watkinson, 2013).

The Mogok metamorphic belt occurs along the western margin of the Shan Plateau and to the west of the Slate Belt extending up to ~1500 km (Mitchell et al., 2007; Searle et al., 2007) (Fig. 1b). This belt consists of schist, gneiss, granulite, quartzite, marble, calc-silicate rock, and migmatite, with various granitoid intrusions. These metamorphic rocks were regionally metamorphosed under medium- to high-grade amphibolite facies and localized granulite facies conditions (e.g., Barley et al., 2003; Mitchell et al., 2007; Searle et al., 2007; Yonemura et al., 2013; Maw Maw Win et al., 2016). Numerous geochronological studies have focused on the Mogok metamorphic belt and the adjacent areas (Bertrand et al., 1999, 2001; Barley et al., 2003; Garnier et al., 2006; Searle et al., 2007; Mitchell et al., 2012; Maw Maw Win et al., 2016). The timing of the main metamorphic event has been estimated as Paleogene and younger (Bertrand et al., 1999, 2001; Searle et al., 2007; Yonemura et al., 2016). Numerous geochronological studies have focused on the Mogok metamorphic belt and the adjacent areas (Bertrand et al., 1999, 2001; Barley et al., 2003; Garnier et al., 2006; Searle et al., 2007; Mitchell et al., 2012; Maw Maw Win et al., 2016). The timing of the main metamorphic event has been estimated as Paleogene and younger (Bertrand et al., 1999, 2001; Searle et al., 2007; Yonemura et al., 2013; Maw Maw Win et al., 2016).

The Onzon and Thabeikkyin areas, the focus of our study, are located 100 km north of the Mandalay region and correspond to the middle segment of the Mogok metamorphic belt (Fig. 2). This segment forms a zone with an average width of over 40 km that trends ENE and corresponds to the middle segment of the Mogok metamorphic belt and the adjacent areas (Bertrand et al., 1999, 2001; Barley et al., 2003; Garnier et al., 2006; Searle et al., 2007; Mitchell et al., 2012; Maw Maw Win et al., 2016). The timing of the main metamorphic event has been estimated as Paleogene and younger (Bertrand et al., 1999, 2001; Searle et al., 2007; Yonemura et al., 2013; Maw Maw Win et al., 2016).

One sample of a cordierite-bearing garnet-biotite paragneiss (OZ04b) and three cordierite-free garnet-biotite paragneisses (OZ07, OZ10, and KB02) were selected for detailed mineralogical analysis. Samples labeled OZ are from the Onzon area, and sample KB02 is from the Thabeikkyin area (Fig. 2b). Table 1 lists the mineral assemblages of these samples. Abbreviations for minerals and end-members are after Whitney and Evans (2010). In the cordierite-bearing paragneiss sample (OZ04b), garnet, cordierite, biotite, sillimanite, plagioclase, K-feldspar, and quartz occur as major phases. Accessory phases are rutile, ilmenite, graphite and zircon. Subhedral garnet porphyroblasts, 3–5 mm in diameter, contain fine grains of sillimanite, cordierite, and quartz inclusions (Fig. 4a). Cracks are developed around the cordierite inclusion, which is partially replaced by pinite (Fig. 4b). The matrix cordierite commonly contains symplectic grains of quartz and rare spinel (Figs. 4c–4e). Isolated grains of cordierite are rarely observed in the matrix. Biotite grains are reddish brown and occur as an inclusion phase in garnet and cordierite as well as in the matrix. Rutile grains occur as both an isolated phase and an intergrowth phase with ilmenite (Fig. 5). Most of these rutile grains are prismatic or rounded, and their average widths are 20–200 µm.

PETROGRAPHY

Granulite facies paragneisses of Mogok metamorphic belt

One sample of a cordierite-bearing garnet-biotite paragneiss (OZ04b) and three cordierite-free garnet-biotite paragneisses (OZ07, OZ10, and KB02) were selected for detailed mineralogical analysis. Samples labeled OZ are from the Onzon area, and sample KB02 is from the Thabeikkyin area (Fig. 2b). Table 1 lists the mineral assemblages of these samples. Abbreviations for minerals and end-members are after Whitney and Evans (2010). In the cordierite-bearing paragneiss sample (OZ04b), garnet, cordierite, biotite, sillimanite, plagioclase, K-feldspar, and quartz occur as major phases. Accessory phases are rutile, ilmenite, graphite and zircon. Subhedral garnet porphyroblasts, 3–5 mm in diameter, contain fine grains of sillimanite, cordierite, and quartz inclusions (Fig. 4a). Cracks are developed around the cordierite inclusion, which is partially replaced by pinite (Fig. 4b). The matrix cordierite commonly contains symplectic grains of quartz and rare spinel (Figs. 4c–4e). Isolated grains of cordierite are rarely observed in the matrix. Biotite grains are reddish brown and occur as an inclusion phase in garnet and cordierite as well as in the matrix. Rutile grains occur as both an isolated phase and an intergrowth phase with ilmenite (Fig. 5). Most of these rutile grains are prismatic or rounded, and their average widths are 20–200 µm.

Common mineral assemblages of cordierite-free paragneisses are garnet, biotite, plagioclase, K-feldspar, quartz, and graphite, with minor ilmenite and zircon. Garnets are 2–4 mm in diameter and occur as subhedral porphyroblasts. In sample OZ10, garnets include numerous grains of acicular sillimanite and quartz (Fig. 4f). Garnet grains in sample OZ07 are partially replaced by muscovite and chlorite. In sample KB02, K-feldspar is more abundant than other samples, and garnet grains are partially replaced by muscovite-rich aggregates along cracks.

ANALYTICAL PROCEDURES

Quantitative analyses and X-ray mapping of the major phases were carried out using a JEOL JXA-8900R (WDS + EDS) and a JCA-733 electron-probe microanalyzer (EPMA) housed at Nagoya University. The accelerating voltage and specimen current for quantitative analyses were kept at 15 kV and 12 nA on the Faraday
cup, respectively. A beam of 5 µm in diameter was used for mica and feldspar analyses, and a beam of 2–3 µm in diameter was used for the analyses of other phases. Well-characterized natural and synthetic phases were employed as standards, including synthetic ZrO$_2$ and KNbO$_3$ for Zr and Nb analyses, respectively. Matrix corrections were performed using the $\alpha$-factor table of Kato (2005). The interferences of TiK$\beta$ and VK$\alpha$ were corrected for the quantitative analysis of V in rutile. The detection limits (2$\sigma$ level) of Nb and Zr in rutile were 230–260 and 270–310 ppm, respectively. Analytical errors in Nb and Zr measurements (1$\sigma$ level) varied slightly depending on their concentrations, and were 190–270 and 200–310 ppm, respectively. The Fe$^{3+}$/Fe$^{2+}$ values of spinel were estimated assuming the stoichiometry and charge valance. Representative chemical data of the silicate and oxide phases are listed in Tables 2, 3, and 4.

Whole-rock compositions used for the $P$-$T$ pseudo-

Figure 2. Geological maps of (a) the middle segment of the Mogok metamorphic belt and the adjacent regions (part of Fig. 2 of Mitchell et al., 2012), and (b) detailed geological map of the Thabeikkyin and Onzon areas with sample localities (Myint Lwin Thein et al., 1990).
section analysis were determined by an X-ray fluorescence spectrometer (XRF; Rigaku ZSX Primus II equipped with a Rh X-ray tube operated at 60 kV and 50 mA) at Nagoya University. The glass beads were prepared by fusing mixtures of powder samples and Li2B4O7 in a weight ratio of 0.7:6.0. Sedimentary and igneous rock reference samples issued by the Geological Survey of Japan (GSJ) were used for the calibration.

MINERAL CHEMISTRY

Garnet

Garnet grains are the solid solution of the almandine–pyrope series with low contents of grossular and spessartine. They show a zoning pattern with more pyrope-rich homogeneous cores and narrow (50–200 µm in width), more almandine-rich rims (Figs. 6 and 7). The compositional range of the homogeneous core of garnet grains in sample OZ04b is Alm54–63Prp33–42Sps1–2Grs2–3. The compositions of garnets adjacent to cordierite inclusions are locally pyrope-poor and almandine-rich (Figs. 6 and 7). In sample OZ04b, fine-grained garnet grains also occur as symplectitic aggregates with biotite and quartz, and have higher almandine and lower pyrope contents (Alm69–70Prp25–26Sps2Grs3) than do the porphyroblastic garnets. The homogeneous cores of the garnet grains in samples OZ07 (Alm59–69Prp35–36Sps2Grs3–4) and OZ10 (Alm56–64Prp31–40Sps1–2Grs2–3) are slightly lower in pyrope and higher in almandine contents than those of sample OZ04b. Garnet in sample KB02 has relatively low pyrope and high grossular contents (Alm59–64Prp29–33

<table>
<thead>
<tr>
<th>Sample</th>
<th>Grt</th>
<th>Bt</th>
<th>Pl</th>
<th>Kfs</th>
<th>Qz</th>
<th>Ilm</th>
<th>Rt</th>
<th>Crd</th>
<th>Spl</th>
<th>Sil</th>
<th>Mnz</th>
<th>Zr</th>
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<tbody>
<tr>
<td>OZ04b</td>
<td>b, ic</td>
<td>+</td>
<td>b</td>
<td>+</td>
<td>b</td>
<td>+</td>
<td>b</td>
<td>s</td>
<td>b, ic</td>
<td>+</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZ07</td>
<td>+</td>
<td>+</td>
<td>+</td>
<td>b</td>
<td>+</td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>OZ10</td>
<td>b</td>
<td>+</td>
<td>+</td>
<td>b</td>
<td>+</td>
<td></td>
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<td></td>
<td></td>
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</tr>
<tr>
<td>KB02</td>
<td>b</td>
<td>+</td>
<td>+</td>
<td>b</td>
<td>+</td>
<td></td>
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</table>

+, present in the matrix; b, present both in the matrix and as the inclusion in garnet; ig, inclusion in garnet; ic, inclusion in cordierite; s, present as symplectic phase in cordierite. Apatite and graphite occur as accessory phases in all samples.

Table 1. Mineral assemblages of paragneisses from the Mogok metamorphic belt

Figure 3. Photographs of Onzon-area outcrops of (a) garnet paragneiss, (b) cordierite-bearing garnet-biotite paragneiss (sample OZ04b), and (c) garnet paragneiss showing leucosome layers, and a Thabeikkyin-area outcrop of (d) garnet paragneiss (sample KB02). Color version is available online from http://doi.org/10.2465/jmps.160526.
Figure 4. (a) AlKα X-ray image map of the garnet porphyroblast in sample OZ04b containing numerous inclusions of cordierite, sillimanite, and quartz; back-scattered electron images of cordierite (b) included in garnet, (c) and (d) coexisting with symplectitic spinel; (e) symplectitic quartz in sample OZ04b; and (f) a garnet porphyroblast in sample OZ10. Abbreviations: Pnt, pinite. Color version is available online from http://doi.org/10.2465/jmps.160526.
Sps3Grs4–7) compared to the Onzon samples. Both TiO2 and Cr2O3 contents are less than 0.1 wt%.

**Cordierite**

Cordierite grains have chemical compositions close to the idealized formula, except that they contain small amounts of Na2O (up to 0.3 wt%), MnO <0.1 wt%, ZnO <0.2 wt%, Cr2O3 <0.1 wt%, CaO <0.1 wt%, and K2O <0.05 wt%. The $X_{Mg} = Mg/(Mg + Fe^{2+})$ value and Na content are slightly variable depending on their modes of occurrence (Fig. 8). Cordierite inclusions in garnet are slightly more magnesian ($X_{Mg} = 0.82–0.85$) and sodic (up to 0.05 atoms per formula unit (apfu) for $O = 18$) than other types of cordierite grains occurring in the matrix. In general, cordierite grains in the matrix are slightly lower in Na (<0.03 apfu). The $X_{Mg}$ values of matrix cordierite grains are slightly variable with their modes of occurrence (Fig. 8); isolated grains (0.76–0.79), grains in contact with the garnet (0.77–0.83), symplectitic grains with quartz (0.77–0.82), and symplectitic grains with spinel (0.78–0.81).

**Biotite**

Biotite commonly occurs as isolated grains in the matrix and as an aggregate around a garnet porphyroblast. The isolated biotite grains of the Onzon samples are a Ti-rich variety (up to TiO2 = 7.0 wt% and 0.39 apfu for $O = 11$) with $X_{Mg} = 0.52–0.66$ (Fig. 9). The biotite aggregates around garnet are poorer in Ti (0.18–0.33 apfu) and slightly higher in $X_{Mg}$ values (0.59–0.72) compared to the coexisting isolated biotite grains (Fig. 9). In sample OZ04b, biotite grains included in cordierite have similar $X_{Mg}$ values (0.58–0.64) to the matrix biotite (0.59–0.66). In contrast, biotite grains included in garnet show higher $X_{Mg}$ values (0.64–0.73) than other biotite grains. Biotite grains that occur as inclusions in cordierite and garnet have higher Ti contents, 0.26–0.36 and 0.29–0.37 apfu,
respectively, similar to the matrix biotite. BaO and ZnO contents of the Onzon biotite grains reach 0.5 and 0.4 wt%, respectively. Fluorine contents are up to 1.2 wt% and chlorine contents less than 0.1 wt%.

Biotite grains in sample KB02 are lower in Ti contents and $X_{\text{Mg}}$ values than those of the Onzon samples; the Ti contents and $X_{\text{Mg}}$ values are 0.20–0.28 apfu and 0.44–0.55 for isolated biotite grains, and 0.22–0.26 apfu and 0.50–0.52 for biotite grains around garnets, respectively. The BaO and ZnO contents of biotite grains are both less than 0.2 wt%. Fluorine and chlorine contents reach 0.9 and 0.3 wt%, respectively.

**Feldspars**

Plagioclase grains are homogeneous and albite-rich with minor orthoclase components: OZ04b, An$_{26±1}$Ab$_{71±1}$Or$_{3±1}$; OZ07, An$_{30±1}$Ab$_{66±1}$Or$_{2±1}$; OZ10, An$_{21±1}$Ab$_{76±1}$Or$_{2±1}$; and KB02, Ab$_{88±1}$Or$_{11±1}$. The albite and BaO contents of K-feldspar grains reach Ab$_{14}$ and 1.4 wt%, Ab$_{35}$ and 0.5 wt%, and Ab$_{15}$ and 0.6 wt% in samples OZ04b, OZ10 and KB02, respectively.

**Rutile and other minerals**

Rutile occurs as isolated and intergrowth grains with il-
menite in the matrix of the cordierite-bearing sample OZ04b (Fig. 5). These grains have near end-member compositions with minor amounts of iron (<0.4 wt% as FeO) and Cr$_2$O$_3$ (<0.5 wt%). V$_2$O$_3$ contents reach 1.1 wt% in some isolated grains (Table 4). Niobium and Zr contents differ slightly between the isolated and intergrowth grains (Fig. 10). The isolated grains have homogeneous compositions of 2470 ± 510 ppm Zr and 2590 ± 390 ppm Nb. In contrast, the three intergrowth grains have compositions that are slightly variable on a grain-to-grain basis: 1250 ± 710 ppm Zr and 2590 ± 390 ppm Nb in grain 01 (Fig. 5b), 4720 ± 870 ppm Zr and 3780 ± 460 ppm Nb in grain 02 (Figs. 5b and 5c), and 4230 ± 380 ppm Zr and 4330 ± 1200 ppm Nb in grain 03 (Fig. 5d). Tantalum contents are below the detection limit of 810 ppm (2σ level).

The chemical composition of ilmenite is close to an end-member composition: the maximum Cr$_2$O$_3$, V$_2$O$_3$, MnO, and MgO contents are 0.14, 0.33, 0.31, and 1.3 wt%, respectively. Spinel compositions are close to spinel-hercynite solid solution with $X_{Mg} = 0.34$–$0.35$, $Y_{Al} = Al/(Al + Fe^{3+} + Cr + V) = 0.98$, Cr$_2$O$_3 = 0.2$–$0.5$ wt%, V$_2$O$_3 = 0.1$–$0.2$ wt%, and ZnO = 2.0–2.4 wt%. Total iron content as Fe$_2$O$_3$ in sillimanite reaches 0.7 and 1.1 wt% in samples OZ04b and OZ10, respectively.

**METAMORPHIC PRESSURE AND TEMPERATURE**

**Empirical geothermometers**

**Ti-in-biotite geothermometer.** Biotite grains in all analyzed samples have high Ti contents and coexist with rutile and ilmenite in the cordierite-bearing sample (OZ04b), and with ilmenite in other samples (OZ07, OZ10, and KB02) (Table 1). The matrix biotite grains usually contain these Ti-rich phases as inclusions, and are considered to have formed under Ti-saturated envi-

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**Table 3.** Representative analyses of major minerals in the cordierite-free garnet-biotite paragneiss from the Thabeikkyin area

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>KB02</th>
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<tbody>
<tr>
<td><strong>Mineral</strong></td>
<td><strong>Core</strong></td>
</tr>
<tr>
<td>SiO$_2$</td>
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<tr>
<td>TiO$_2$</td>
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<tr>
<td>Al$_2$O$_3$</td>
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<tr>
<td>Cr$_2$O$_3$</td>
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<td>FeO$^*$</td>
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<td>MnO</td>
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<tr>
<td>MgO</td>
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<tr>
<td>ZnO</td>
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</tr>
<tr>
<td>BaO</td>
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</tr>
<tr>
<td>CaO</td>
<td>2.14</td>
</tr>
<tr>
<td>Na$_2$O</td>
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<tr>
<td>K$_2$O</td>
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</tr>
<tr>
<td>F</td>
<td>0.61</td>
</tr>
<tr>
<td>Cl</td>
<td>0.07</td>
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<tr>
<td>O = F, Cl</td>
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</tr>
<tr>
<td>Total</td>
<td>100.41</td>
</tr>
<tr>
<td>Oxygen</td>
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<td>Si</td>
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<tr>
<td>Ti</td>
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<tr>
<td>Cr</td>
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<tr>
<td>Fe$^{3+*}$</td>
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<tr>
<td>Mn</td>
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</tr>
<tr>
<td>Mg</td>
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<td>Zn</td>
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<td>Ba</td>
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<td>Ca</td>
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<td>Na</td>
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<tr>
<td>Total</td>
<td>8.016</td>
</tr>
<tr>
<td>$X_{Mg}$</td>
<td>0.34</td>
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</table>

* Total iron as FeO. $X_{Mg} = Mg/(Mg + Fe)$. 

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Granulite facies paragneisses of Mogok metamorphic belt
Compositional and textural characteristics of biotite enable temperature estimation with the Ti–in-biotite geothermometer (Fig. 9; e.g., Henry and Guidotti, 2002). Matrix biotite grains of the Onzon samples yield rather constant temperatures of 780–820 °C (Figs. 9a–9c) and 670–770 °C (at 0.8 GPa) using the calibrations of Henry et al. (2005) and Wu and Chen (2015), respectively. Matrix biotite grains in the Thabeikkyin sample record temperatures of 710–760 °C (Henry et al., 2005) (Fig. 9d) and 650–720 °C (Wu and Chen, 2015). However, biotite grains around garnets show slightly lower temperatures than the matrix grains: 720–800 °C (Onzon) and 720–750 °C (Thabeikkyin) with the calibration of Henry et al. (2005), and 630–730 °C (Onzon) and 590–690 °C (Thabeikkyin) with the calibration of Wu and Chen (2015).

Biotite inclusions in garnets of an Onzon sample (OZ04b) show higher temperatures than the matrix biotite (Fig. 9a). Garnets around the biotite inclusions have re-equilibrated during the retrograde stage and locally show lower $X_{Mg}$ values than the peak metamorphic compositions. This re-equilibration process likely yielded increased $X_{Mg}$ values of biotite inclusions without any change in Ti contents, which resulted in disequilibrium state between the biotite inclusions and Ti-rich phases in matrix. Consequently, the higher temperature conditions shown by biotite inclusions are apparent and probably do not give any meaningful data for the discussion of metamorphic condition. Some biotite inclusions in cordierite show slightly lower $X_{Mg}$ values than the matrix biotite grains. The Mg–Fe partitioning coefficient between cordierite and biotite, $(Mg/Fe)_{Crd}/(Mg/Fe)_{Bt}$, is greater than unity (e.g., Thompson, 1976), and thus the $X_{Mg}$ values of biotite equilibrated with cordierite are inferred to have decreased with decreasing temperature. The slightly lower $X_{Mg}$ values of some biotite grains in cordierite were likely caused by local re-equilibration during the retrograde stage.

**Zr-in-rutile geothermometer.** Rutile grains in sample OZ04b coexist with zircon as matrix phases and were employed for temperature estimation using the Zr-in-rutile geothermometer (e.g., Zack et al., 2004). Rutile grains occur as isolated grains and as intergrowth grains with ilmenite (Fig. 5). The isolated rutile grains have fairly constant Zr contents of 2470 ± 510 ppm (Fig. 10) and indicate a temperature of 845 ± 25 °C (at 0.8 GPa) using the calibrations of Tomkins et al. (2007). However, on a grain-to-grain basis, intergrowth rutile grains show vari-

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* Total iron as FeO. 
Iso, isolated grain; Igw, intergrowth grain.
able temperature conditions from $760 \pm 70$ to $930 \pm 25 \, ^\circ C$.

**Conventional geothermobarometers**

The Mogok paragneisses contain the common mineral assemblage of garnet–biotite–plagioclase–quartz, and their equilibrium conditions were estimated using a conventional garnet–biotite Mg–Fe exchange geothermometer and garnet–biotite–plagioclase–quartz (GBPQ) net-transfer reaction. In sample OZ04b, sillimanite occurs as a matrix phase; therefore, the garnet–plagioclase–sillimanite–quartz (GPSQ) equilibria calibrated using the dataset of Holland and Powell (1998) were also used to estimate metamorphic conditions. Although many calibrations have been published for the garnet–biotite geothermometer, two proposed by Bhattacharya et al. (1992) and Holdaway (2000) that consider the influence of Al and Ti substitutions in biotite on the temperature estimation were employed in the present study.

Pressure–temperature conditions were estimated us-
ing homogeneous garnet core, isolated matrix biotite, and An–rich plagioclase core. The estimated \(P/T\) conditions for OZ04b, OZ07, OZ10 and KB02 were 0.65–0.73 GPa/810–860 °C, 0.60–0.64 GPa/820–860 °C, 0.67–0.71 GPa/800–830 °C, and 0.75–0.79 GPa/830–860 °C, respectively (Fig. 11a). The cordierite–bearing paragneiss sample (OZ04b) contains cordierite in addition to the common assemblage of garnet–sillimanite–biotite–plagioclase–quartz. Therefore, the Mg–Fe exchange reaction between garnet and cordierite and the following net-transfer reaction were also used for \(P/T\) estimations (Nichols et al., 1992):

\[
2 (\text{Mg,Fe})_3\text{Al}_2\text{Si}_3\text{O}_{12} + 4 \text{Al}_2\text{SiO}_3 \quad \text{garnet} \\
+ 5 \text{SiO}_2 = 3 (\text{Mg,Fe})_2\text{Al}_4\text{Si}_5\text{O}_{18} \quad (\text{I}). \\
\text{quartz} \quad \text{cordierite}
\]

Two sets of garnet and cordierite compositions were employed for these calculations: (i) those of homogeneous garnet core and matrix cordierite free of symplectitic intergrowth, and (ii) those of garnet rim and adjacent cordierite. Sillimanite occurs as inclusion grains in garnet and both of these types of cordierite, which suggests that both sets of garnet-cordierite coexist with sillimanite. These two sets of garnet-cordierite compositions yield \(P/T\) estimations of about 0.65 GPa/820 °C and 0.40 GPa/620 °C, respectively (Fig. 11b).

**\(P/T\) pseudosection**

The \(P/T\) pseudosection (Fig. 12) was computed for the whole-rock composition of a cordierite-bearing paragneiss sample (OZ04b). The pseudosection was made by forward modeling using Perple_X_6.7.4 (Connolly, 1990) employing the internally consistent data set of Holland and Powell (1998). The calculation was performed in the system MnO–TiO₂–K₂O–Na₂O–CaO–FeO–MgO–Al₂O₃–SiO₂–H₂O (MnTiKNCFMASH), and fluid of H₂O is considered to be excess phase. The whole-rock composition (in wt%) of sample OZ04b is SiO₂ = 69.99, TiO₂ = 0.84, Al₂O₃ = 12.38, total iron as FeO = 5.70, MnO = 0.09, MgO = 3.48, CaO = 1.34, Na₂O = 1.76, K₂O = 2.35, P₂O₅ = 0.07, Loss on ignition = 0.77. The sample contains apatite in thin section, implying that P₂O₅ was sequestered in apatite. Thus, we corrected CaO content as 1.25 wt% for the MnTiKNCFMASH phase diagram analysis. Selected solution models for the relevant minerals in the pseudosection analysis are listed in figure caption.

Isochemical phase diagram (Fig. 12) suggests that (1) solidus temperature of sample OZ04b slightly decreases from 680 °C at 0.6 GPa to 660 °C at 1.0 GPa and (2) cordierite becomes stabilized under partial melting conditions. The inclusion assemblage of cordierite + biotite + sillimanite in garnet has narrow \(P/T\) stability range of 0.68 GPa/700 °C–0.78 GPa/730 °C. In contrast, the sillimanite–orthopyroxene-free matrix assemblage of cordierite + biotite + garnet + rutile is stable under slightly higher temperature conditions of 0.68–0.78 GPa/710–765 °C. In sample OZ04b, spinel occurs as symplectitic aggregate with cordierite. However, the
calculated phase diagram does not show any spinel-bearing fields. Stability of spinel strongly depends on Zn content (e.g., Nichols et al., 1992), and the Zn-free system employed for the phase diagram might contract the stability field of spinel.

DISCUSSION

Progressive metamorphism

Conventional geothermobarometers yield granulite facies equilibria around 0.60–0.79 GPa/800–860 °C for the Mogok metamorphic rocks of the Onzon and Thabeikkyin areas. The $P$-$T$ pseudosection analysis of the inclusion phases in garnet and matrix assemblage suggests prograde path from 700–730 to 710–765 °C at 0.68–0.78 GPa, though the temperature conditions are slightly lower than those estimated by the conventional methods. Temperature conditions estimated with empirical geothermometers are rather variable. The Ti-in-biotite geothermometer gives temperature conditions of 650–820 °C, and the higher temperature estimations within this range are consistent with those estimated with the conventional geothermometer. The Zr-in-rutile geothermometer results

Figure 9. $X_{Mg}$-Ti diagrams showing compositional variation of biotite in paragneiss samples from the Onzon and Thabeikkyin areas, and various isotherms based on the Ti-in-biotite geothermometer (Henry et al., 2005).
Figure 10. Compositional variation diagram of Zr and Nb (ppm) of rutile in a cordierite-bearing garnet-biotite paragneiss (OZ04b) and their histograms. Temperature calibration follows Tomkins et al. (2007). Analytical errors are 1σ level.

Figure 11. Pressure-temperature estimations of paragneisses from the Onzon and Thabeikkyin areas in the (a) garnet-biotite-plagioclase-quartz system and (b) garnet-cordierite-sillimanite-quartz system. Abbreviations: GB, garnet-biotite geothermometer; GASP, garnet-plagioclase-alumina silicate-quartz system; GBPQ, garnet-biotite-plagioclase-quartz system; B92, Bhattacharya et al. (1992); H00, Holdaway (2000); HP98, calibrated using the dataset of Holland and Powell (1998); N92, Nicholls et al. (1992); W04, Wu et al. (2004); Fe-crd, Fe-cordierite. Phase equilibria of aluminum silicates are from Pattison (2001). Sets (1) and (2) denote two compositional datasets of garnet and matrix phases.
indicate that equilibrium temperatures are variable with the mode of occurrence of rutile. The isolated rutile grains yield fairly constant estimates of temperature conditions (845 ± 26 °C) that are consistent with the inferred granulite facies metamorphism. In contrast, the intergrowth rutile grains record different temperature conditions on a grain-to-grain basis, and some grains (grains 02 and 03) yield distinctly higher temperature estimates (915 ± 10 and 930 ± 25 °C) than those estimated with empirical methods and those recorded in isolated rutile grains. A possible interpretation of these higher temperature estimates is that back-diffusion of Zr may have metastably enriched the Zr content of the remaining rutile during the replacement of rutile by ilmenite. However, this scenario cannot adequately explain the observation that an intergrowth rutile (grain 01) has lower Zr content than the isolated rutile grains. Another possible interpretation is that the intergrowth grains were detrital in origin, which may explain the variable compositional ranges of the intergrowth rutile grains. Although we have little information for understanding the origin of the intergrowth rutile grains, the isolated rutile grains may represent Zr amounts equilibrated with zircon at the peak metamorphic stage. Figure 13 summarizes P-T estimations of this study together with previously published estimations. High-temperature estimations over 800 °C have been reported from a garnet-orthopyroxene granulate in the Mogok area (0.65–0.87 GPa/800–950 °C; Yonemura et al., 2013) and a spinel-bearing paragneiss and the associated rocks in the Onzon area (0.77–0.84 GPa/780–850 °C; Ye Kyaw Thu et al., 2016). Maw Maw Win et al. (2016) also reported high-temperature paragneisses (0.6–1.0 GPa/780–850 °C) from the Sagaing ridge of the Mandalay region. These results suggest widespread distribution of upper-amphibolite to granulite facies metamorphic rocks in the middle segment of the Mogok belt. Searle et al. (2007), however, reported lower P-T conditions of 0.48 GPa/625 °C and 0.49 GPa/680 °C for a highly deformed sillimanite gneiss from the Kyaukse area and an augen gneiss from the Kyanigan area, respectively (Fig. 13). Kyanite was reported from schistose rocks of the Kyanigan area (Searle et al., 2007). The garnet grains in the Kyaukse and Kyanigan samples are distinctly higher in spessartine and lower in pyrope contents (Alm72–77Prp9–12Sps9–15Grs2–5) than those in the Onzon–Thabeikkyin (Alm53–66Prp44–51Sps1–3Grs2–7: Ye Kyaw Thu et al., 2016; This study) and Sagaing (Alm54–78Prp14–41Sps1–4Grs3–7: Maw Maw Win et al., 2016) samples. The Kyaukse sample contains muscovite as a peak metamorphic phase. These facts support the lower estimations of temperature for the peak metamorphic stage of the Kyaukse and Kyanigan samples. The higher-temperature metamorphic rocks (Yonemura et al., 2013; Maw Maw Win et al., 2016; Ye Kyaw Thu et al., 2016; this study) were all collected from outcrops at a mappable scale. However, the lower-temperature Kyaukse and Kyanigan samples were collected from small outcrops and quarries in areas covered with Quaternary sediments to the east of the Mo-
gok metamorphic rocks of the Sagaing ridge in the Mandalay region. Although the geological relationships between the higher- and lower-temperature metamorphic rocks are not well understood, similar U–Th–Pb zircon and monazite ages of Late Eocene–Oligocene were reported from both the Kyaukse and Kyanigan areas (e.g., Searle et al., 2007) and the Sagaing ridge (Maw Maw Win et al., 2016). Kyanite-bearing schistose rocks occur in the Minwun ridge on the western side of the Sagaing high-grade rocks bordered by the Sagaing Fault (Fig. 13). Maw Maw Win (2016) estimated P–T conditions of these rocks as 0.41–0.68 GPa/570–640 °C, which is similar to those of the Kyaukse and Kyanigan samples. The Minwun samples contain prograde muscovite and garnet grains with similar chemical compositions (Alm68–72Prp13Sp57–13Grs4–6) to those in the Kyaukse and Kyanigan samples. She inferred that the Minwun low-temperature metamorphic rocks were originally from the southern part of the Mogok metamorphic belt and may have been displaced by dextral movement of the Sagaing Fault to the present position and juxtaposed with the high-grade metamorphic rocks of the Sagaing ridge. These results suggest that (1) both high- and low-temperature metamorphic rocks were simultaneously formed by the Mogok metamorphism, (2) metamorphic rocks of the kyanite-sillimanite facies series occur below Quaternary sediments in the middle segment of the Mogok belt, and (3) metamorphic grade seems to progressively increase perpendicular to the extensional direction of the Mogok belt: e.g., from east to west towards the Sagaing Fault in the Mandalay and surrounding regions and from south to north in the Mogok region.

**P–T trajectory**

Critical mineral assemblages for the discussion of the P–T evolution of the studied samples are cordierite, sillimanite, and quartz included in garnet (Fig. 4a) and cordierite-spinel symplectite with sillimanite in the matrix (Figs. 4c and 4d). Several melt-related reactions were proposed to explain formation of cordierite under the high-grade amphibolite and granulite facies conditions. Cenki et al. (2002) studied migmatitic cordierite gneisses from southern Pan–African India and showed that many melt-producing and subsequent melt-consuming reactions had progressed extensively. They also provided reaction formulae to explain the formations of cordierite + spinel and cordierite + quartz as follows:

\[
\text{Bt + Grt (+ Sil + Pl)} = \text{Spl + Crd + Melt (+ Ilm)} \tag{4}\]

\[
\text{Grt + Melt} = \text{Crd + Bt + Oz} \tag{7a}\]

According to petrogenetic grid for the K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O system proposed by Cenki et al. (2002), the cordierite grains containing symplectic quartz and spinel may have formed during a nearly adiabatic pressure decrease from 0.9 to 0.5 GPa at around 800–900 °C. However, the cordierite-quartz symplectite in sample OZ04b contains a small amount of biotite (Figs. 4c and 4e), and the formation of this texture might be explained by the reaction (7a).
Figure 14 shows reaction relationships in the model ZnFMAS system as well as the possible P–T trajectory for sample OZ04b. Another possible reaction explaining the textural relationships among sillimanite in garnet and cordierite–spinel symplectite is as follows:

\[
\begin{align*}
2 \text{Al}_2\text{SiO}_5 + (\text{Mg,Fe})_3\text{Al}_2\text{Si}_3\text{O}_{12} & \quad \text{sillimanite garnet} \\
= (\text{Mg,Fe})_2\text{Al}_4\text{Si}_5\text{O}_{18} + (\text{Mg,Fe})\text{Al}_2\text{O}_4 & \quad \text{cordierite spinel} \\
\end{align*}
\]

(II) [Qz] in Fig. 14.

In a quartz-excess system, however, the reaction (I) (reaction [Spl] in Fig. 14) is stable instead of the spinel–bearing reaction (II). Therefore, if the quartz-free reaction II formed the cordierite–spinel symplectite, it had progressed in a local SiO\textsubscript{2}–undersaturated domain. An alternative reaction to form both of these symplectites is:

\[
\begin{align*}
2 \text{Al}_2\text{SiO}_5 + 5 \text{SiO}_2 & \quad \text{spinel quartz}
\end{align*}
\]

\[
\begin{align*}
= (\text{Mg,Fe})_2\text{Al}_4\text{Si}_5\text{O}_{18} & \quad \text{cordierite} \\
\end{align*}
\]

(III) [Grt, Sil] in Fig. 14.

In this case, the reactants of spinel and quartz were considered to have been governed by the reaction such as:

\[
\begin{align*}
2 \text{Al}_2\text{SiO}_5 + (\text{Mg,Fe})_3\text{Al}_2\text{Si}_3\text{O}_{12} & \quad \text{sillimanite garnet} \\
= 2 (\text{Mg,Fe})\text{Al}_2\text{O}_4 + 5 \text{SiO}_2 & \quad \text{spinel quartz} \\
\end{align*}
\]

(IV) [Crd] in Fig. 14.

The P–T pseudosection of the sample OZ04b shows no spinel-bearing field (Fig. 12), and there is no textural evidence implying the presence of a spinel + quartz assemblage before the formation of the cordierite–spinel symplectite in the sample OZ04b. However, this assemblage was reported from a paragneiss collected from an outcrop close to the site of sample OZ04b in the Onzon area (OZO5a: Ye Kyaw Thu et al., 2016). Thus, metamorphic P–T conditions of the Onzon area probably reached into those of the spinel + quartz stability field. If the spinel + quartz assemblage was stable before the formations of the symplectitic aggregates of cordierite–spinel and cordierite–quartz, they probably represent local equilibrium in spinel- and quartz-dominated local domains, respectively.

The melt-bearing reactions (4) and (7a) (Cenki et al., 2002) suggested a nearly adiabatic pressure decrease for the formations of the cordierite–spinel and cordierite–quartz symplectites. The reactions II [Qz] and III [Grt, Sil] both have relatively strong pressure dependence (Fig. 14) (Seifert and Schumacher, 1986; Nichols et al.,...
1992). We have information to discuss formation process of the sympletects. However, the sympletectic cordierite grains probably formed during the early stage of exhumation under upper-amphibolite or granulite facies conditions. Considering the equilibrium conditions of 0.3 GPa/620 °C and estimated equilibrium biotite + plagioclase + quartz in paragneisses from the Mogok metamorphic belt during exhumation (0.55 GPa/640 °C) was reported from a cordierite-bearing paragneiss (OZ05a; Ye Kyaw Thu et al., 2016). Maw Maw Win et al. (2016) and Enami et al. (2017) also reported sympletectic aggregates of garnet + biotite + plagioclase + quartz in paragneiss from the Sagaing ridge (Fig. 13) and estimated equilibrium conditions of 0.3–0.5 GPa/600–680 °C. Considering the CHIME (chemical Th–U–total Pb isochron method) monazite age in the paragneiss, Maw Maw Win et al. (2016) interpreted that the formation of sympletectite was related to fluid activity at ~ 28 Ma. Therefore, extensive recrystallization with infiltration of metamorphic fluid likely occurred throughout the middle segment of the Mogok metamorphic belt during exhumation in the late Oligocene.

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SUPPLEMENTARY MATERIALS

Color version of Figures 3, 4, and 6 is available online from http://doi.org/10.2465/jmps.160526.

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