Amphibole in UHP eclogite from the Sulu region, eastern China

Satoshi YAMASAKI*, Daisuke NAKAMURA* and Takao HIRAJIMA**

*Graduate School of Natural Science and Technology, Okayama University, Okayama 700-8530, Japan
**Division of Earth and Planetary Sciences, Graduate School of Science, Kyoto University, Kyoto 606-8502, Japan

We examined the stability of amphibole in ultrahigh-pressure (UHP) conditions of about 620–750 °C, 2.9–3.7 GPa based on the mineralogy of UHP eclogites collected from the Qinglongshan and Jianchang areas in Donghai County of the Sulu region. The samples studied are mainly composed of garnet, omphacite, amphibole, rutile, kyanite, quartz, epidote, and phengite. Amphibole can be classified into three types based on its mode of occurrence; Amp I, inclusions in garnet and epidote; Amp II, the matrix phase; Amp III, small (~ 10–30 µm) grains developed along grain boundaries. Amp I from Qinglongshan, mainly barroisite–taramite, is slightly poorer in NaB (Na contents in B site < 1.2 a.p.f.u.) than Amp II. Amp II is barroisite with a mostly homogeneous composition in each grain except for the margins of some grains (NaB = 1.1–1.4 a.p.f.u.; Si = 7.0–7.5 a.p.f.u.) and shows the highest NaB values among the types. Amp III, mainly pargasite–taramite, shows clearly lower NaB (= 0.3–1.15 a.p.f.u.) than that of Amp II. On the other hand, Amp II and Amp III from Jianchang are nybøite and taramite, respectively, and the sum of the (NaA + NaB) values from the Jianchang amphibole is distinctly greater than that of the Qinglongshan amphibole. Polycrystalline quartz pseudomorphs after coesite and Amp I with barroisite compositions were found within a large epidote grain of the Qinglongshan samples. These petrological data suggest the following evolutionary history. Amp I with relatively low NaB values crystalized and was involved with other mineral grains during the compression of the eclogite, and then barroisitic Amp II occurred as a stable phase during the peak–P stage. Finally, Amp III with much lower NaB values was formed during the decompression stage. Thus, we concluded that a Ca–Na amphibole (Amp II) can be stably present in an eclogite even in the UHP conditions. In addition, significant amounts of F (up to 1 wt%) were detected from the barroisitic amphibole in the Qinglongshan samples. At the least, some eclogites from the studied areas may have undergone metasomatic infiltration of an F-bearing fluid in the UHP conditions.

Keywords: Amphibole, Eclogite, Ultrahigh-pressure, Fluorine, Sulu region

INTRODUCTION

Amphibole is a hydrous mineral containing about 2 weight percent (wt%) of H₂O and the examination of its stability is important in providing some constraints on H₂O-transport processes in the convergent plate boundary. For example, the release of H₂O by the decomposition of hydrous minerals from descending material and the upward flow of H₂O into the mantle wedge can trigger melt generation and earthquakes in a subduction zone (e.g., Tatsumi, 1989; Ulmer, 2001; Grove et al., 2006; Abers et al., 2013). High-pressure and high-temperature synthetic experiments have shown that Ca-rich amphibole should be unstable in mid-ocean-ridge-basalt (MORB) systems in ultrahigh-pressure (UHP) conditions (e.g., Liu et al., 1996; Schmidt and Poli, 1998), suggesting that H₂O stored in such a Ca-rich amphibole should be released before the subducting mafic rocks reach the UHP conditions at about 100 km depth, although Ca-rich amphibole is considered to be stable in hydrated ultramafic rocks even in the UHP conditions (e.g., Niida and Green, 1999). Its breakdown can trigger melt generation in a subduction zone (Tatsumi, 1989). On the other hand, natural amphiboles in mafic rocks of a subducting plate are generally sodic and sub-calcic, such as glaucophane/riebeckite and barroisite, at medium depths of a subduction zone, respectively (e.g., Otsubo and Banno, 1990; Vitale Brovarone and Agard, 2013; Hyppolito et al., 2016; Kabir and Takasu, 2016; Kato and Hirajima, 2017). Furthermore, some petrological observations of natural UHP metamorphic rocks suggest that Na-rich amphiboles can be stable even in the UHP conditions, such as glaucophane in the sodic whiteschist of the Dora–Maira Massif, western Italy.
(e.g., Kienast et al., 1991), and nybøite in eclogites from the Western Gneiss Region of Norway and the Sulu region of eastern China (e.g., Smith, 1988; Hirajima et al., 1992). Thus, the stability of hydrous minerals in mafic and ultramafic systems is crucial information for understanding the fluid circulation system in subduction-zone environments. However, reports on the presence of Ca or Ca–Na amphiboles stable in the UHP conditions are limited, as previously mentioned, except only a few reports (e.g., Taguchi et al., 2016).

In this study, we discuss whether the amphibole in eclogites collected from the Qinglongshan area in Donghai County (DC) of the Sulu region (Fig. 1) was stably present in the UHP conditions or not, based on detailed petrological observation, analyses of mineral compositions, and comparison of these data to the Na–amphibole in eclogite from the Jianchang area of DC (Hirajima et al., 1992; Oberti et al., 2003). The mode of the occurrence of amphibole in the Qinglongshan and Jianchang eclogites is similar, but there is a distinct chemical difference between them; i.e., the major amphiboles in the matrix are a Ca–Na type in the Qinglongshan eclogite and an Na type in the Jianchang eclogite. Furthermore, garnet and omphacite in the Jianchang eclogite are richer in Fe and Na compared to those in the Qinglongshan eclogite, respectively. These chemical differences can be explained by the difference in the bulk composition. In addition, fluorine contents of amphibole were found to be relatively high (up to about 1 wt%) in the Qinglongshan eclogite, as was previously noted for the Na–amphibole in the Jianchang eclogite (Oberti et al., 2003). These data support that a Ca–Na amphibole can be stable even in the UHP conditions with a suitable bulk composition. The detailed petrography of the Qinglongshan eclogite is described.

**GEOLOGICAL OUTLINE**

The study area, the Sulu region, is considered to be a regional UHP metamorphic belt formed by the continental collision between the Yangtze and the Sino-Korean cratons (e.g., Hirajima and Nakamura, 2003) and is on the Shandong Peninsula of eastern China extending nearly 500 km from Weihai County (northeastern part) to Donghai County (DC) (southwestern part) (Fig. 1). The Yantai–Qingdao–Wulian (YQW) fault bounds the northern side of the Sulu region, in which coesite–bearing eclogite and garnet peridotite sporadically occur as lenses or blocks enclosed in the country gneiss (e.g., Enami et
Coesite has been discovered not only from the eclogite but also from the country gneiss (e.g., Ye et al., 2000; Liu et al., 2004), supporting the idea that the whole body of the Sulu region has undergone UHP metamorphism; i.e. an in-situ model (e.g., Wallis et al., 1997).

Protolith ages of the gneiss and eclogite are about 700–800 Ma or older, and metamorphic ages 220–240 Ma (e.g., Tang et al., 2008; Hacker et al., 2009).

The studied eclogite samples were collected from the Qinglongshan area in DC of the Sulu region (Fig. 1b). In this area, the occurrence of talc in eclogite has been reported (Mattinson et al., 2004) but we did not find talc in our samples. Instead, the collected samples include relatively large amphibole grains in the matrix, although they are pristine eclogite in which middle- or low-P decompression stage minerals are rarely present. For comparison, we also re-analyzed Na-amphibole-bearing UHP eclogite samples collected from the Jianchang area in DC and previously described by Hirajima et al. (1992).

PETROGRAPHY

We observed eight fist-sized rock samples (samples names are Q1, Q2, Q3, Q4, Q6, Q7, Q8, and Q9) from Qinglongshan, and additional analyses were performed on four samples from Jianchang (samples names are J3, J4, J10, and J13). The mineral abbreviations used in this paper follow those of Whitney and Evans (2010).

We first classified the samples from Qinglongshan into three types; equigranular eclogite mainly consisting of garnet and omphacite (referred to as type Q1234) (Fig. 2a), samples including poikiloblasts of epidote and amphibole and fine-grained (~ 0.2 mm in diameter) garnet (referred to as type Q689) (Fig. 2b), and a sample including poikiloblastic epidote and amphibole, in which medium-sized grains (~ 1–2 mm in diameter) of garnet and omphacite are included (referred to as type Q7) (Fig. 2c).

Type Q1234

These samples mainly comprise relatively coarse-grained garnet and omphacite, and most of the grains are larger than 1 mm in diameter. Other main constituent minerals are amphibole, epidote, quartz, phengite, kyanite, and rutile. Fine-grained (<1–3 µm) symplectite of clinopyroxene + plagioclase partially replaces omphacite, but such replacement by symplectite is very limited (Fig. 2a).

Amphibole inclusions, which are likely to form during the prograde stage, are present in many garnet grains (Figs. 3a and 3b). Some of the amphibole inclusions show subhedral shapes (Fig. 3a) but others are anhedral. In addition, garnet grains include quartz pseudomorphs after coesite, and amphibole inclusions and such pseudomorphs are included in the same garnet grains (Fig. 3b). In this type (Q1234), medium-sized (~ 1–2 mm in diam-
Figure 3. Photomicrographs of the studied samples from Qinglongshan. (a) An amphibole inclusion in garnet in sample Q2. Small amphibole grains (Amp III) developed along the grain boundary. Plane-polarized light (PPL). (b) Amphibole inclusions in garnet with quartz pseudomorph after coesite in sample Q3. PPL. (c) Amphibole in the matrix in sample Q1. PPL. (d) An amphibole inclusion in an epidote grain in sample Q8. PPL. (e) Quartz pseudomorphs after coesite in same epidote grain as (d). PPL. (f) Crossed polars view of (e). (g) Amphibole in the matrix in sample Q6. PPL. (h) Amphibole in the matrix including garnet with quartz pseudomorph after coesite in sample Q7. PPL. Grt, garnet; Omp, omphacite; Amp, amphibole; Qz, quartz; Coe(ps), quartz pseudomorph after coesite; Ep, epidote; Ky, kyanite.
Grains of amphibole can be observed in the matrix (Fig. 3c). Some amphibole grains in the matrix show massive shapes, whereas others show irregular shapes developing through other mineral grains. Along the margin of the amphibole in the matrix, symplectite consisting of amphibole + plagioclase grains with sizes of about 10–20 µm develop in some places. Small (~10–30 µm) amphibole grains are present along the grain boundaries between garnet and omphacite in some locations (Fig. 3a). Many amphibole grains are devoid of inclusions, however, some grains include quartz and garnet.

**Type Q689**

In these samples, garnet and omphacite grains are relatively small, and their grain sizes are about 0.2 mm in diameter. On the other hand, epidote and amphibole grains develop as poikiloblasts (Fig. 2b), and several grains are longer than 10 mm along the long axis. The constituent minerals are the same as those in the type Q1234. Garnet, kyanite, omphacite, rutile, quartz, and amphibole inclusions are present in the epidote poikiloblasts (e.g., Figs. 3d–3f). An epidote grain includes polycrystalline quartz with radial cracks (Figs. 3e and 3f), indicating that coesite was trapped in the epidote grain during its growth and the coesite was subsequently transformed to quartz during the decompression. Small amphibole grains are also included in this epidote grain (Figs. 3d and 4). Amphibole poikiloblasts in the matrix include garnet, kyanite, omphacite, rutile, and quartz (Fig. 3g). Thus, epidote and amphibole in the matrix contain abundant inclusions, whereas garnet and omphacite in these eclogites are mostly devoid of inclusions.

**Type Q7**

This eclogite also shows a poikiloblastic texture as was observed in type Q689. Epidote and amphibole grains are as large (about 10 mm along the long axis) as those in type Q689, whereas the garnet and omphacite are not as small as those in type Q689 (Fig. 2c). The mineral assemblage of type Q7 is the same as that in types Q1234 and Q689. Fine-grained (<1–3 µm) symplectite of clinopyroxene + plagioclase is also partially developed in this sample.

In this sample, a polycrystalline quartz pseudomorph after coesite is also present within a garnet grain that is surrounded by amphibole (Fig. 3h). Phengite flake cleavages show directions subparallel to each other and defining the foliation. Epidote and amphibole grains in the matrix, however, developed in random directions and cut the foliation (Fig. 2c). Long-axis directions of omphacite inclusions in the epidote are mostly the same as those of omphacite in the matrix, suggesting that these epidote and amphibole grains formed simultaneously with and/or after the formation of the eclogite-facies minerals.

**Jianchang**

Amphibole in the matrix of eclogite from this locality has been confirmed to be Na–amphibole with the chemical compositions of nybøite. It has been interpreted to be stable in the UHP conditions by a previous work (Hirajima et al., 1992).

Fine-grained (<1–3 µm) symplectite of clinopyroxene + plagioclase is more abundant in the analyzed four thin sections than that in the samples from Qinglongshan. The mineral assemblages are garnet + omphacite + amphibole + epidote + quartz + phengite + rutile ± kyanite. Kyanite grains do not occur in samples J10 and J13, whereas samples J3 and J4 comprise kyanite. Sample J3 has compositional bandings that can be defined by the difference in the modal amounts of garnet, omphacite, and epidote. Sample J4 is relatively rich in large bluish amphibole grains in the matrix, but the fine-grained symplectite are remarkably developed. Samples J10 and J13 include small amounts of bluish amphibole in the matrix and do not show clear foliation.

**MINERALOGY**

**Analytical procedures**

Amphibole, garnet, omphacite, and epidote compositions in samples from Qinglongshan and amphibole, garnet, and omphacite compositions in samples from Jianchang were analyzed using an electron probe microanalyzer (EPMA) at Okayama University.

Quantitative analyses of major elements were per-
formed using wavelength-dispersive X-ray spectrometry at Okayama University (JEOL JXA-8230). A 15-kV accelerating potential and a 20-nA beam current were employed in the analyses. The standards were as follows: quartz (SiO$_2$) for Si, rutile (TiO$_2$) for Ti, corundum (Al$_2$O$_3$) for Al, hematite (Fe$_2$O$_3$) for Fe, manganosite (MnO) for Mn, periclase (MgO) for Mg, wollastonite (CaSiO$_3$) for Ca, jadeite (NaAlSi$_2$O$_6$) for Na, potassium titylan phosphate (KTiOPO$_4$) for K, tugtupite (Na$_4$AlBeSi$_4$O$_{12}$Cl) for Cl, fluorite (CaF$_2$) for F, and strontium titanate (SrTiO$_3$) for Sr. The representative compositions of amphibole, garnet, and omphacite in the thin sections are listed in Tables 1, 2, and 3, respectively.

**Amphibole**

In this study, we classified amphibole into three types, based on the mode of occurrence; 1) amphibole inclusions in garnet or epidote (referred to as Amp I; e.g., Figs. 3a, 3b, and 3d), 2) relatively large (>1 mm in diameter) amphibole grains in the matrix (referred to as Amp II; e.g., Figs. 3c, 3g, and 3h), and 3) small (~ 10–30 µm) amphibole grains developed along grain boundaries (referred to as Amp III; e.g., Fig. 3a). The chemical formulae of amphibole were calculated by assuming that only Ca, Na, and K occupy the A and B sites, and the total of Fe$^{total}$, Mg, Mn, Al, Ti, and Si is equal to 13. Fe$^{3+}$ contents were calculated to create a total positive electric charge of 46, based on Si + Ti + Al + Fe + Mn + Mg = 13. In addition, Ca was treated as present only in the B site, and K was assumed to be only in the A site. Thus, Na contents in the B site (referred to as NaB) were calculated as NaB = 2 – Ca a.p.f.u., and Na contents in the A site (referred to as NaA) as NaA = Na$^{total}$ – NaB a.p.f.u. All analyzed data of amphiboles show that the K contents are less than 0.09 a.p.f.u.

Comparing the NaB values of amphibole from type Q1234, Amp I is slightly poorer in NaB (= 0.5–1.1 a.p.f.u.) content than the majority of Amp II (NaB = 1.1–1.3 a.p.f.u.) (Fig. 5). In addition, in amphiboles from types Q689 and Q7, the NaB content of Amp I tends to be less (NaB = 0.8–1.2 a.p.f.u., except for one data point) than that of Amp II (NaB = 1.1–1.4 a.p.f.u., except for some data). The amphibole inclusions show compositions of barroisite or taramite (according to Leake et al., 1997), whereas the amphibole in the matrix is mainly barroisite. The NaB values of Amp III from type Q1234 are less than 1.15 a.p.f.u. (Fig. 5) and compositions of Amp III are mainly pargasite or taramite. Compositional ranges of these amphiboles show negative correlations between NaB and NaA, and the NaA values increase with decreasing the NaB values (Fig. 5).

Differences in compositions of amphiboles among types Q1234, Q689, and Q7 are present but not distinct. The NaB values of Amp II from type Q7 are the highest (NaB = 1.3–1.4 a.p.f.u., except for some data) among them, and those from types Q1234 and Q689 are slightly less (NaB = 1.1–1.3 a.p.f.u. for type Q1234; NaB = 1.2–1.3 a.p.f.u. for type Q689, excluding some data) (Fig. 5). Amp II from all types of the Qinglongshan samples is classified as barroisite, and Amp I in epidote that includes quartz pseudomorphs after coesite (Figs. 3d and 4) is also barroisite.

Si contents of amphiboles from Qinglongshan range from 5.5 to 7.5 a.p.f.u (Fig. 6), and a negative correlation is present between Si and the NaA values. Si contents of Amp I range from 5.5 to 7.3 a.p.f.u., and Amp II tends to show higher Si contents (= 7.0–7.5 a.p.f.u., except for some data) than those of Amp I and Amp III. The Fe$^{2+}$/Fe$^{3+}$ ratios of amphiboles from Qinglongshan are lower than 0.45, and those of the majority of Amp II are about 0.1 (Fig. 6). The Al contents of amphiboles from Qinglongshan range from 1.5 to 3.6 a.p.f.u., and Amp II shows lower Al values (~ 1.5–2.0 a.p.f.u.) than those of Amp I and Amp III. The TiO$_2$ contents are 0.05–0.20 wt% in Amp II from Qinglongshan.

We also analyzed Cl and F contents of amphiboles from the Qinglongshan samples. Although the Cl contents are very low (<0.018 a.p.f.u.; <0.080 wt%), significant amounts of F (up to 0.45 a.p.f.u.; 1.0 wt%) were detected (Fig. 7). Particularly, Amp II from types Q689 and Q7 contains distinctly greater amounts of F (= 0.12–0.45 a.p.f.u.; 0.26–1.02 wt%) than those from Q1234 (F < 0.12 a.p.f.u.; <0.27 wt%) (Fig. 7). Amphibole (K richterite) in kimberlites from Kimberley, South Africa, also includes significant amounts of F (= 0.29–0.69 wt%), whereas amphibole (kaersutite) in island-arc igneous rocks from Japan contains small amounts of F (= 0.09–0.17 wt%) (Aoki and Kanisawa, 1979). In contrast, the F contents of Na amphibole in lawsonite blueschist in northwest Turkey are significantly lower (<0.04 wt%), althoughapatite with high-F contents (~ 3 wt%) is present in the blueschist (Pagé et al., 2016). F contents of lawsonite in the blueschist are also not as high (~ 0.04 wt%). Thus, the F contents of Amp II from types Q689 and Q7 are as high as or higher than those of amphibole in kimberlites. In addition, Na-amphibole from the Jian-chang eclogite has been reported to have a large amount of F (= 2.58 wt%; 1.16 a.p.f.u.) (Oberti et al., 2003).

On the chemical zoning of Amp II from types Q1234, the NaB values are high in the core parts (NaB ~ 1.2 a.p.f.u.) but decrease near the margins (NaB ~ 0.8) (Fig. 8a). On the other hand, Amp II from type Q689 has a homogenous core (NaB ~ 1.2 a.p.f.u.) and displays a
Table 1. Representative chemical compositions of amphibole in the studied samples from Qinglongshan and Jianchang

|   | Q2 | Q3 | Q4 | Q5 | Q6 | Q7 | Q8 |
|---|----|----|----|----|----|----|----|
| SiO₂ | 47.41 | 48.06 | 53.23 | 53.07 | 49.79 | 40.63 | 49.53 |
| TiO₂ | 0.40 | 0.25 | 0.05 | 0.11 | 0.14 | 0.01 | 0.12 |
| Al₂O₃ | 13.43 | 12.73 | 9.91 | 10.96 | 11.80 | 18.37 | 12.59 |
| FeO | 9.15 | 10.53 | 7.96 | 7.77 | 9.84 | 13.86 | 8.45 |
| MnO | 0.04 | 0.11 | 0.00 | 0.10 | 0.04 | 0.14 | 0.05 |
| MgO | 13.58 | 13.61 | 15.16 | 14.33 | 12.97 | 9.37 | 14.50 |
| CaO | 8.44 | 7.70 | 5.97 | 5.15 | 6.92 | 8.83 | 7.20 |
| Na₂O | 4.15 | 4.49 | 5.06 | 5.18 | 4.62 | 4.59 | 5.09 |
| K₂O | 0.06 | 0.02 | 0.10 | 0.15 | 0.06 | 0.03 | 0.41 |
| F | 0.25 | 0.08 | 0.21 | 0.23 | 0.17 | 0.17 | 0.10 |
| Cl | 0.01 | 0.01 | 0.00 | 0.00 | 0.00 | 0.01 | 0.00 |
| -O=FeCl | 0.11 | 0.04 | 0.09 | 0.10 | 0.07 | 0.07 | 0.43 |
| Total | 96.81 | 97.55 | 97.30 | 96.95 | 96.28 | 96.04 | 97.94 |

Fe³⁺ contents were calculated to create a total positive electric charge = 46, based on Si + Ti + Al + Fe + Mn + Mg = 13.
slight increase in the NaB toward the margin (NaB ~ 1.3 a.p.f.u.) (Fig. 8b).

Amphibole compositions from the Jianchang eclogites are plotted in Figure 9 for comparison. Total Na con-

| SiO₂ | 39.69 | 39.62 | 39.89 | 40.48 | 39.17 | 39.12 | 39.09 | 40.24 | 39.62 | 39.25 |
| TiO₂ | 0.05  | 0.07  | 0.01  | 0.01  | 0.06  | 0.04  | 0.03  | 0.03  | 0.03  | 0.01  |
| Al₂O₃ | 22.08 | 22.41 | 22.17 | 22.36 | 22.02 | 22.20 | 23.82 | 21.98 | 22.56 | 21.94 |
| FeO   | 20.27 | 18.67 | 20.39 | 20.52 | 20.48 | 20.90 | 22.15 | 20.88 | 21.84 | 21.14 |
| MnO   | 0.44  | 0.39  | 0.51  | 0.66  | 0.56  | 0.59  | 0.69  | 0.52  | 0.78  | 0.74  |
| MgO   | 10.37 | 10.91 | 10.28 | 10.87 | 9.65  | 9.48  | 8.40  | 9.50  | 8.55  | 9.51  |
| CaO   | 6.57  | 7.06  | 6.71  | 7.33  | 7.41  | 6.96  | 6.63  | 8.37  | 6.82  | 6.48  |
| Na₂O | 0.02  | 0.03  | 0.01  | 0.02  | 0.00  | 0.07  | 0.01  | 0.03  | 0.04  | 0.04  |
| K₂O  | 0.01  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.00  | 0.02  | 0.00  | 0.00  |
| Total | 99.50 | 99.16 | 99.97 | 102.24 | 99.37 | 99.29 | 100.88 | 101.55 | 100.23 | 99.11 |

Table 2. Representative chemical compositions of garnet in the studied samples from Qinglongshan and Jianchang

Formulæ were calculated based on total oxygen = 12.
tents of Amp II (NaA + NaB) from Jianchang are distinctly higher (Na > 1.8 a.p.f.u.) than those from Qinglongshan (Na < 1.6 a.p.f.u.), excluding some data (Fig. 9). Amp I and Amp III from Jianchang also show higher Na (~ 1.4–2.2 a.p.f.u.) and NaA contents (~ 1.0) than those from Qinglongshan (Na < 1.6 a.p.f.u.; NaA < 0.9 a.p.f.u.) (Fig. 9). Amp II from sample J4 is classified as ny böïte, as already described by Hirajima et al. (1992). Thus, the com-

### Table 3. Representative chemical compositions of omphacite in the studied samples from Qinglongshan and Jianchang

|        | Q1     | Q2     | Q3     | Q4     | Q5     | Q6     | Q7     | Q8     | Q9     | Q10    | J3     | J4     | J5     | J6     | J7     | J8     | J9     | J10    | J11    | J12    | J13    |
|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| SiO₂   | 56.63  | 56.91  | 57.30  | 57.45  | 57.23  | 55.97  | 56.41  | 56.30  | 56.26  | 56.26  | 99.44  | 100.35 | 101.13 | 100.79 | 99.54  | 99.84  | 100.41 | 99.79  | 100.28 |
| TiO₂   | 0.09   | 0.06   | 0.06   | 0.03   | 0.04   | 0.04   | 0.05   | 0.10   | 0.07   | 0.08   | 2.00   | 1.989  | 1.987  | 1.984  | 2.00   | 1.987  | 1.984  | 2.00   | 1.987  | 1.983  |        |
| Al₂O₃  | 14.94  | 14.33  | 15.13  | 15.03  | 15.19  | 12.50  | 12.24  | 11.99  | 12.77  | 12.75  | 11.05  | 11.13  | 11.32  | 12.06  | 11.94  | 9.68   | 9.61   | 9.88   | 9.71   |        |
| FeO    | 7.22   | 8.22   | 7.16   | 8.59   | 8.20   | 7.42   | 7.81   | 8.85   | 7.56   | 7.67   |        |        |        |        |        |        |        |        |        |        |        |
| MnO    | 0.00   | 0.01   | 0.00   | 0.00   | 0.02   | 0.00   | 0.01   | 0.04   | 0.00   | 0.06   |        |        |        |        |        |        |        |        |        |        |        |
| MgO    | 3.45   | 3.35   | 3.37   | 4.02   | 3.06   | 5.22   | 4.98   | 4.87   | 4.87   | 5.07   |        |        |        |        |        |        |        |        |        |        |        |
| CaO    | 6.26   | 6.39   | 6.01   | 4.95   | 5.11   | 8.70   | 8.73   | 8.37   | 8.65   | 8.68   |        |        |        |        |        |        |        |        |        |        |        |
| Na₂O   | 11.05  | 11.13  | 11.32  | 12.06  | 11.94  | 9.68   | 9.61   | 9.88   | 9.61   | 9.71   |        |        |        |        |        |        |        |        |        |        |        |
| K      | 0.00   | 0.01   | 0.00   | 0.00   | 0.00   | 0.00   | 0.01   | 0.01   | 0.01   | 0.01   |        |        |        |        |        |        |        |        |        |        |        |
| Total  | 99.64  | 100.41 | 100.35 | 101.13 | 100.79 | 99.54  | 99.84  | 100.41 | 99.79  | 100.28 |        |        |        |        |        |        |        |        |        |        |        |

Fe³⁺ contents were calculated to make O = 6 based on total cations = 4.
positions of amphiboles are clearly different between Qinglongshan and Jianchang, but those in both areas show negative correlations between NaA and NaB.

Garnet

Garnet in the samples from Qinglongshan mainly comprises almandine \( X_{\text{alm}} = \frac{\text{Fe}}{(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})} \sim 0.4-0.6 \), pyrope \( X_{\text{prp}} = \frac{\text{Mg}}{(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})} \sim 0.25-0.45 \), and grossular \( X_{\text{grs}} = \frac{\text{Ca}}{(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})} \sim 0.1-0.3 \) components with a small amount of spessartine \( X_{\text{sps}} = \frac{\text{Mn}}{(\text{Fe} + \text{Mg} + \text{Mn} + \text{Ca})} < 0.03 \) component (Fig. 10). Garnet in type Q1234 is slightly poorer in the grossular component \( X_{\text{grs}} = 0.10-0.20 \) than that in type Q689 \( X_{\text{grs}} = 0.16-0.30 \), and garnet in type Q7 shows intermediate compositions between types Q1234 and Q689 \( X_{\text{grs}} = 0.15-0.24 \) (Fig. 10). Thus, Ca contents of garnet likely reflect the textural difference among types Q1234, Q7, and Q689; i.e. Ca-rich minerals such as epidote are large and abundant in type Q689, whereas these are small and not as abundant in type Q1234 (Fig. 2), sug-
suggesting the samples of type Q689 are richer in Ca than those of type Q1234 in the bulk composition. The difference in garnet compositions between the rock types was probably controlled by the difference in the bulk composition. In type Q689, garnet grains occur both as a matrix phase and as inclusions in the epidote and amphibole and show no clear difference in composition between them. Some garnet grains show chemical zoning, although many are nearly homogeneous. For example, garnet grains including Amp I display clear chemical zoning. Near the amphibole inclusion, Mg and Fe contents decrease and increase toward the inclusion, respectively (Fig. 11).

Garnet from Jianchang tends to show higher Fe contents ($X_{\text{alm}} \approx 0.5 - 0.6$) than that from Qinglongshan ($X_{\text{alm}} \approx 0.4 - 0.5$) (Fig. 10). Grossular contents of garnet from Jianchang, tend to show higher Fe contents than that from Qinglongshan. Garnet from Jianchang tends to show higher Fe contents ($X_{\text{alm}} \approx 0.5 - 0.6$) than that from Qinglongshan ($X_{\text{alm}} \approx 0.4 - 0.5$) (Fig. 10). Grossular contents of garnet from Jianchang, tend to show higher Fe contents than that from Qinglongshan.

**Figure 9.** Amphibole compositions in the Jianchang (J) eclogite. Amp I is amphibole inclusions in other minerals, Amp II is amphibole in the matrix, and Amp III is small amphibole grains developed along grain boundaries. The NaA and NaB are Na contents in the A and B sites, respectively. Gln, glaucophane; Nyb, nybøite; others are the same as in Figure 5.

**Figure 10.** Grossular (Grs) - pyrope (Prp) - almandine (Alm) triangle diagram showing garnet compositions in the studied samples. Types Q1234, Q689, and Q7 are samples from Qinglongshan area. Samples J3, J4, J10, and J13 are samples from Jianchang area.
Omphacite

The jadeite contents \( X_{jd} = Al - 2(2 - Si - Ti) \), based on total cations = 4, of omphacite grains in the Qinglongshan eclogite vary from 0.33 to 0.49, and there is a positive correlation between \( X_{jd} \) and Si contents. The Fe\(^{2+}\)/\((Fe^{2+} + Mg)\) ratios of these are <0.36, when Fe\(^{3+}\) contents are stoichiometrically calculated as making O = 6 based on total cations = 4. Their ratios increase with increasing \( X_{jd} \) (Fig. 12). Similar positive correlations between \( X_{jd} \) and Fe\(^{2+}\)/\((Fe^{2+} + Mg)\) have been noted for other UHP eclogites from the Sulu region (Hirajima, 1996; Nakamura et al., 2015). There is no distinct compositional difference in omphacite among the rock types in Qinglongshan (Fig. 12). The omphacite grains in the Qinglongshan eclogite are present not only in the matrix but also as inclusions in amphibole and epidote in types Q689 and Q7, but no distinct difference was detected between the matrix grains and the inclusions (Fig. 12).

Omphacite in the Jianchang eclogite has clearly higher jadeite contents \( X_{jd} = 0.47-0.65 \) than that in the Qinglongshan eclogite (Fig. 12). In addition, omphacite in the kyanite-bearing samples (J3 and J4) is richer in the jadeite component \( X_{jd} = 0.55-0.65 \) than that in the kyanite-free samples (J10 and J13) \( X_{jd} = 0.47-0.57 \). The Fe\(^{2+}\)/\((Fe^{2+} + Mg)\) ratios of omphacite in the kyanite-bearing samples are <0.5, and those in the kyanite-free samples are <0.35 (Fig. 12).

Epidote

In the all samples from Qinglongshan, euhedral or subhedral epidote grains are present in the matrix. Sizes are about 2.5 mm in type Q1234 and up to 20 mm along the long axis in types Q689 and Q7. Garnet, omphacite, quartz, amphibole, kyanite, rutile, and phengite are involved as inclusions in epidote. There is a clear negative correlation between Al and Fe\(^{3+}\) contents, and epidote in types Q689 and Q7 is richer in Fe\(^{3+}\) \([Fe^{3+}/(Fe^{3+} + Al) = 0.17-0.25]\) than that in Q1234 \([Fe^{3+}/(Fe^{3+} + Al) = 0.13-0.18]\). Sr contents in epidote were also analyzed, and there is no distinct difference in the Sr contents between the rock types. The SrO contents range from 0.2 to 1.4 wt%, which are nearly the same values reported for epidote in eclogite from DC by Nagasaki and Enami (1998).

DISCUSSION

Pressure-temperature estimation

Here, we estimate pressure-temperature \((P-T)\) conditions during the peak-\(P\) stage of the Qinglongshan eclogite, by utilizing the garnet-clinopyroxene geothermometer (Powell, 1985; Nakamura, 2009) and the garnet-clinopyroxene-kyanite-\(SiO_2\) phase geobarometer. This geobarometer is based on a reaction: pyrope + grossular + 2 coesite (or quartz) = 3 diopside + 2 kyanite. The enthalpies, entropies, and volumes of the mineral endmembers were calculated from the data set by Holland and Powell (1998) and the activities of pyrope, grossular, and diopside follow the models proposed by Nakamura and Ban-no (1997) in which the ordering energy of omphacite is...
also considered. Activities of kyanite and coesite are set to be equal to 1.

For this $P$–$T$ estimation, we selected compositional data of garnet and omphacite pairs by the following criteria. Some garnet grains in the studied samples contain amphibole inclusions in their cores (e.g., Figs. 3a and 3b), and a garnet grain shows an core poorer in Mg near the amphibole inclusion than the margin; i.e. the pyrope and almandine contents decrease and increase toward the amphibole inclusion, respectively (Fig. 11). The chemical compositions during compression and the heating stage have possibly been preserved in the core of the garnet. Therefore, we did not use the compositions of the core or the mantle parts of the garnet. On the other hand, most omphacite grains do not show distinct chemical zonation; hence, we used rim–rim compositional pairs of garnet and omphacite grains which are directly in contact with each other. Jadeite contents of omphacite significantly vary from 0.33 to 0.50, and the $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg})$ ratio of omphacite increases with increasing jadeite contents (Fig. 12). This positive correlation is consistent with that reported by Hirajima (1996) in the Yangkou area of the Sulu region, but such a correlation is not necessarily because of the non-ideality of the clinopyroxene solid-solution. Nakamura et al. (2015) noted that the use of omphacite with $\text{Ca} + \text{Na} + \text{K}$ contents less than 1.00 a.p.f.u. provides excessively high temperatures. The reason for such a temperature deviation is not clear but can be explained at least for $\text{Fe}^{2+}$-poor omphacite. For example, $\text{Fe}^{2+}$ contents of omphacite in the studied eclogite are about 0.04–0.07 a.p.f.u. (Table 3). If $\text{Ca} + \text{Na} + \text{K}$ contents are as low as 0.97 a.p.f.u. and $\text{Fe}^{2+}$ is preferentially accommodated into the M2 site, $\text{Fe}^{2+}$ contents in the M1 site decrease to 0.01–0.04 a.p.f.u. from the aforementioned values. Thus, a small amount of excess accommodation of $\text{Fe}^{2+}$ into the M2 site of omphacite with low-$\text{Fe}^{2+}$ contents can significantly affect $\text{Fe}^{2+}$–$\text{Mg}$ distribution coefficients between garnet and omphacite. However, more detailed discussion on this matter is beyond the scope of this study, and we have ignored data of omphacite with $\text{Ca} + \text{Na} + \text{K}$ contents clearly less than 1.00 a.p.f.u. to estimate $P$–$T$ conditions here. In addition, data of omphacite with $\text{Ca} + \text{Na} + \text{K}$ contents greater than 1.00 are also ignored here. The data of garnet and omphacite adopted for this estimation are listed in Tables 2 and 3, respectively.

The obtained $P$–$T$ conditions are about 640–720 °C, 3.2–3.4 GPa for type Q1234; 620–750 °C, 2.9–3.7 GPa for type Q689; and 700 °C, 3.2 GPa for type Q7 (Fig. 13). These estimated values are within the $P$–$T$ range estimated for type Q689, and the obtained pressures correspond to those of the coesite-stability field. These estimated data on $P$–$T$ conditions support the idea that epidote and amphibole grains grew in the UHP conditions. Epidote and amphibole develop as remarkably large grains in types Q689 and Q7 (Fig. 2). Garnet and omphacite grains are present both in the matrix and as inclusions in the epidote and amphibole (Figs. 3e–3h, and 4), and their compositions are nearly the same between the matrix grains and the inclusions (e.g., Fig. 12). If the large epidote and amphibole formed in the quartz-stability field during decompression, the original high-$P$ compositions of garnet and omphacite would

![Figure 13. Estimated pressure-temperature ($P$-$T$) conditions of the Qinglongshan samples. Bold lines are $P$-$T$ curves obtained from the garnet-clinopyroxene-kyanite-coesite geobarometer: the thermodynamic data-set is from Holland and Powell (1998) and activity models are from Nakamura and Banno (1997). Solid and broken lines are results from the garnet-clinopyroxene geothermometer formulated by Nakamura (2009) and Powell (1985), respectively. The open circles are $P$-$T$ conditions calculated from each chemical composition pair of garnet and omphacite. The diamond (Dia) and graphite (Gr) transformation line and the coesite (Coe) and quartz (Qz) transformation curve were calculated using the data-set of Holland and Powell (1998).]
have been significantly modified to low-\(P\) compositions and we could not obtain such UHP conditions from their compositions. Even if the obtained \(P-T\) conditions do not reflect the peak-\(P\) stage, the equilibrated \(P-T\) conditions would have been within the UHP field.

**Calculated stability of amphibole**

A phase relationship \((P-T)\) pseudosection was calculated by using a fixed bulk composition in the \(K_2O-Na_2O-CaO-FeO-MgO-Al_2O_3-SiO_2-H_2O-O_2\) system. TiO\(_2\), MnO, and P\(_2\)O\(_5\) were assumed to be present only in rutile, garnet, and apatite; hence, these components were excluded from the system. Fluid was treated to be an excess pure H\(_2\)O phase. Amphibole is stable in the UHP conditions, but its stability field is restricted to low-\(T\) conditions. Mineral abbreviations follow those of Whitney and Evans (2010).

Amphiboles are 0.34–0.52 and 0.57–0.82, respectively (Fig. 15), and these compositions can be classified as glaucohane or magnesioriebeckite (Leake et al., 1997) and never barroisite. If so, it is questionable why the prograde inclusions in the studied samples (Amp I) do not have such Na-rich compositions.

**Comparison to Jianchang**

Distinct differences in amphibole, garnet, and clinopyroxene in terms of chemical compositions were recognized between the Jianchang and Qinglongshan samples. Amphibole and clinopyroxene in the Jianchang samples are clearly richer in Na than those in the Qinglongshan samples. For amphibole, the Qinglongshan samples contain Ca-Na amphibole such as barroisite, whereas the Jianchang samples include Na-amphibole such as nybøite (Fig. 9). For clinopyroxene, the jadeite contents are significantly higher in the Jianchang samples than those in the Qinglongshan samples (Fig. 12).

Thus, the total Na contents of amphibole are remarkably different between the Qinglongshan and Jianchang samples, but both areas show negative correlations between NaA and NaB contents in amphibole.
bole in the matrix (Amp II) displays higher the NaB values and lower NaA values than those of Amp I and Amp III (Fig. 9). Thus, the whole compositional range of amphibole in the Jianchang samples does not overlap that of the Qinglongshan samples. These data imply that the compositions of amphibole in the Qinglongshan samples (Ca–Na amphibole) were not created by chemical modification of Na–amphibole such as nypöite in the Jianchang samples. The differences in mineral chemistry between the Qinglongshan and Jianchang mainly reflect the difference in the bulk composition between the two areas; Na contents in the Jianchang samples are probably higher than those in the Qinglongshan samples.

Stability of amphibole

The NaB values in amphibole generally increase with increasing pressure, based on synthetic experimental data (e.g., Poli, 1993; Ernst and Liu, 1998). An amphibole grain in the matrix (Amp II) of sample Q6 shows a slight increase in NaB toward the margin (Fig. 8b), suggesting that at least this grain did not grow during decompression. In addition, the Amp II in type Q689 involves garnet and omphacite as inclusions. Garnet and omphacite in the matrix have the same compositions as the inclusions in amphibole and epidote. Therefore, the timing of chemical equilibrium of garnet and omphacite in the matrix should have been nearly simultaneous with that of the growth of the Amp II and epidote at least in the type Q689 samples. As estimated in the previous section, coesite-stable UHP conditions (620–750 °C, 2.9–3.7 GPa) were obtained for the Qinglongshan samples (Fig. 13). Thus, Amp II should have been stably present in the UHP conditions during the peak-P metamorphic stage. Furthermore, quartz pseudomorphs after coesite were newly found in an epidote grain and some garnet grains, and amphibole inclusions were also found in the same epidote and garnet grains (Figs. 3b, 3d, 3e, 3f, and 4). Coesite in the matrix should be easily transformed to quartz in the quartz-stability field; hence, the presence of pseudomorphs after coesite in epidote and garnet indicates that at least the epidote and garnet grains grew in the coesite-stability field and trapped the matrix coesite as inclusions. Determining from the locations of the amphibole inclusions (Fig. 4), the amphibole and coesite inclusions should have been simultaneously trapped into the host minerals. Such amphibole inclusions (Amp I), which have barroisitic compositions with slightly lower the NaB values than those of Amp II, should have also been present in the coesite-stability field and have undergone UHP metamorphism.

CONCLUSIONS

Based on the petrological data of the amphibole-bearing eclogite from the Sulu region, we concluded that the Ca–Na amphibole (barroisite) can be stably present even in the UHP conditions in eclogite (mafic system). This indicates that H2O contained within amphibole in basaltic material can be carried to depths of more than 100 km below the surface in a subduction zone. However, hydrous minerals such as epidote and amphibole in the studied samples are texturally interpreted to have grown in the UHP conditions. To develop such hydrous minerals, at least some supply of H2O-bearing fluid or breakdown of other hydrous minerals should be necessary, but it is enigmatic how such H2O could be provided or released. In addition, it is also unclear why amphibole and epidote in types Q689 and Q7 are remarkably large whereas the amphibole and epidote in type Q1234 are not as large (Fig. 2). This is possibly related to the difference in F contents of the amphibole between them (Fig. 7). Fluorine contents in basaltic rocks increase with increasing K2O contents (Aoki et al., 1981). K2O contents in basaltic rocks generally tend to increase depending on the depth of the primary magma generation (e.g., Miyashiro, 1974). Therefore, the supply of F from the slab to the mantle wedge probably increases with the
descent of the slab, and the released fluid should become richer in F at greater depths. Thus, infiltration of an F-bearing fluid into types Q698 and Q7 may have occurred in the UHP conditions and promoted the growth of amphibole and epidote in types Q689 and Q7. Such a supply of F may be also related to the formation of F-rich Na-amphibole in Jianchang as reported by Oberti et al. (2003). However, synthetic experiments have indicated that amphibole includes larger amounts of F than those of melt (e.g., Van den Bleeken and Koga, 2015). Therefore, partial melting of eclogite and extraction of melt can result in the residual material being rich in F, although the presence of hydrous phases such as phengite, amphibole, and epidote in the studied eclogite probably does not support such an idea of partial melting and melt extraction. A metasomatic event is the simplest manner in which to explain the enrichment of F in amphibole, but the source material of the F-bearing fluid is still unclear. Thus, a metasomatic event may not be necessary, but fluorine should be an important component to discuss the stabilities of hydrous minerals in the UHP conditions.

ACKNOWLEDGMENTS

The studied samples from Qinglongshan were collected as part of a co-operative project among Kyoto University, Nagoya University, and Hefei University in 1999, and the samples from Jianchang were collected as a co-operative project between Kyoto University and Academia Sinica. We would like to thank the members of these projects, S.R. Wallis, Y. Hiroi, T. Kato, M. Aoya, and M. Enami, and also T. Nozaka and T. Fujise for providing comments on this study; M.F. Kabir and an anonymous reviewer for their critical comments; and M. Enami for his editorial work. Thanks also to T. Fujiwara for preparation of thin sections. This study was financially supported by Grant-in-Aid no. JP16K05610 for DN and JP25257208 for TH.

REFERENCES

Abers, G.A., Nakajima, J., van Keken, P.E., Kita, S. and Hacker, B.R. (2013) Thermal-petrological controls on the location of earthquakes within subducting plates. Earth and Planetary Science Letters, 369-370, 178-187.

Aoki, K. and Kanisawa, S. (1979) Fluorine contents of some hydrous minerals derived from upper mantle and lower crust. Lithos, 12, 167-171.

Aoki, K., Ishikawa, K. and Kanisawa, S. (1981) Fluorine geochemistry of basaltic rocks from continental and oceanic regions and petrogenetic application. Contributions to Mineralogy and Petrology, 76, 53-59.

Connolly, J.A.D. (2009) The geodynamic equation of state: what and how. Geochemistry, Geophysics, Geosystems, 10, Q10014, doi:10.1029/2009GC002540

Diener, J.F.A., Powell, R., White, R.W. andolland, T.J.B. (2007) A new thermodynamic model for clinopyroxene-orthopyroxene and enstatite-diopside-richterite assemblages in the system Na2O-CaO-Al2O3-MgO-SiO2-FeO-O2. Journal of Metamorphic Geology, 25, 631-656.

Enami, M., Zang, Q. and Yin, Y. (1993) High-pressure eclogites in northern Jiangsu - southern Shandong province, eastern China. Journal of Metamorphic Geology, 11, 589-603.

Ernst, W.G. and Liu, J. (1998) Experimental phase-equilibrium study of Al- and Ti-contents of calcic amphibole in MORB—A semi-quantitative thermobarometer. American Mineralogist, 83, 952-969.

Foley, S. (1991) High-pressure stability of the fluorine-hydroxide endmembers of paragosite and kaolinite. Geochimica et Cosmochimica Acta, 55, 2689-2694.

Green, E., Holland, T. and Powell, R. (2007) An order-disorder model for omphacitic pyroxenes in the jadeite-diopside-edenite-acmite, with applications to eclogitic rocks. American Mineralogist, 92, 1101-1109.

Griew, T.L., Chatterjee, N., Parman, S.W. and Méard, E. (2006) The influence of H2O on mantle wedge melting. Earth and Planetary Science Letters, 249, 74-89.

Hacker, B.R., Wallis, S.R., Mcwilliams, M.O. and Gans, P.B. (2009) 40Ar/39Ar Constraints on the tectonic history and architecture of the ultrahigh-pressure Sulu orogeny. Journal of Metamorphic Geology, 27, 827-844.

Hirajima, T. (1996) Effect of jadeite-content on the garnet-clinopyroxene geothermometer for an ultrahigh-pressure eclogite complex. Proceedings of the Japanese Academy, 72B, 208-213.

Hirajima, T., Zhang, R., Li, J. and Cong, B. (1992) Petrology of the Dabie Shan complex. Proceedings of the Japanese Academy, 68B, 298-303.

Hirajima, T. and Nakamura, D. (2003) Review of representative UHPM terranes: The Dabie Shan - Sulu orogen. In Ultrahigh Pressure Metamorphism; EMU Notes in Mineralogy 5 (Carswell, D.A. and Compagnoni, R. Eds.). pp. 508, Eötvös University Press, Budapest, 105-144.

Holland, T.J.B. and Powell, R. (1998) An internally consistent thermodynamic data set for phases of petrological interest. Journal of Metamorphic Geology, 16, 309-343.

Hypolito, T., Angiboust, S., Juliani, C., Glodny, J., Garcia-Casco, A., Calderón, M. and Chopin, C. (2016) Eclogite-, amphibolite- and blueschist-facies rocks from Diego de Almagro Island (Patagonia): Episodic accretion and thermal evolution of the Chilean subduction interface during the Cretaceous. Lithos, 264, 422-440.

Kabir, M.F. and Takasu, A. (2016) Jadeite-garnet glaucophane schists in the Bizan area, Sambagawa metamorphic belt, eastern Shikoku, Japan: significance and extent of eclogite facies metamorphism. Journal of Metamorphic Geology, 34, 893-916.

Kato, R. and Hirajima, T. (2017) Petrology and possible significance of barroisite-bearing metabasite from the Keba Formation in NW Kii Peninsula. Journal of Mineralogical and Petrological Sciences, 112, 40-45.

Kienast, J.R., Lombardo, B., Biino, G. and Pinardon, J.L. (1991) Petrology of high-pressure eclogite rocks from the Brossasco–Isasca Complex, Dora-Maira Massif, Italian Western Alps. Journal of Metamorphic Geology, 9, 19-34.

Leake, B.E., Woolley, A.R., Arps, C.E.S., Birch, W.D., Gilbert,
The amphibole in UHP eclogite

M.C., Grice, J.D., Hawthorne, F.C., Kato, A., Kisch, H., Kri-rovichev, V.G., Linthout, K., Laird, J., Mandarino, J.A., Mar-esch, W.V., Nickel, E.H., Rock, N.M.S., Schmurcher, J.C., Smith, D.C., Stephenson, N.C.N., Ungaretti, L., Whittaker, E.J.W. and Youzhi, G. (1997) Nomenclature of amphiboles: report of the subcommittee on amphiboles of the International Mineralogical Association, commission on new minerals and mineral names. The Canadian Mineralogist, 35, 219-246.

Liu, F., Xu, Z., Liu, J.G. and Song, B. (2004) SHRIMP U-Pb ages of ultrahigh-pressure and retrograde metamorphism of gneisses, south-western Sulu terrane, eastern China. Journal of Metamorphic Geology, 22, 315-326.

Liu, J., Bohlen, S.R. and Ernst, W.G. (1996) Stability of hydrous phases in subducting oceanic crust. Earth and Planetary Science Letters, 143, 161-171.

Mattinson, C.G., Zhang, R.Y., Tsujimori, T. and Liu, J.G. (2004) Epidote-rich talc-kyanite-phengite eclogites, Sulu terrane, eastern China: P-T-fO2 estimates and the significance of the epidote-talc assemblage in eclogite. American Mineralogist, 89, 1772-1783.

Miyashiro, A. (1974) Volcanic rock series in island arcs and active continental margins. American Journal of Science, 274, 321-355.

Nagasaki, A. and Enami, M. (1998) Sr-bearing zoisite and epidote in ultra-high-pressure (UHP) metamorphic rocks from the Su-Lu province, eastern China: an important Sr reservoir under UHP conditions. American Mineralogist, 83, 240-247.

Nakamura, D. (2009) A new formulation of garnet-clinozoisite geothermometer based on accumulation and statistical analysis of a large experimental data set. Journal of Metamorphic Geology, 27, 495-508.

Nakamura, D. and Banno, S. (1997) Thermodynamic modeling of sodic pyroxene solid-solution and its application in a garnet-omphacite-kyanite-coesite geobarometer for UHP metamorphic rocks. Contributions to Mineralogy and Petrology, 130, 93-102.

Nakamura, D., Okada, M. and Hirajima, T. (2015) Effect of clinozoisite composition on Fe-Mg distribution coefficient between garnet and clinopyroxene. Journal of Mineralogical and Petrological Sciences, 110, 82-97.

Niida, K. and Green, D.H. (1999) Stability and chemical composition of pargasitic amphibole in MORB pyrolite under upper mantle conditions. Contributions to Mineralogy and Petrology, 135, 18-40.

Oberti, R., Boioecchi, M. and Smith, D.C. (2003) Fluoronybőite from Jianzhang (Su-Lu, China) and nybőite from Nybó (Nordfjord, Norway): a petrological and crystal-chemical comparison of these two high-pressure amphiboles. Mineralogical Magazine, 67, 769-782.

Otsuki, M. and Banno, S. (1990) Prograde and retrograde metamorphism of hematite-bearing basic schists in the Sanbagawa belt, central Shikoku. Journal of Metamorphic Geology, 8, 425-439.

Pagé, L., Hattori, K., de Hoog, J.C.M. and Okay, A.I. (2016) Halogen (F, Cl, Br, I) behavior in subducting slabs: A study of lawsonite blueschists in western Turkey. Earth and Planetary Science Letters, 442, 133-142.

Poli, S. (1993) The amphibolite-eclogite transformation: an experimental study on basalt. American Journal of Science, 293, 1061-1107.

Powell, R. (1985) Regression diagnostics and robust regression in geothermometer / geobarometer calibration: the garnet-clinozoisite geothermometer revisited. Journal of Metamorphic Geology, 3, 231-243.

Schmidt, M.W. and Poli, S. (1998) Experimentally based water budgets for dehydrating slabs and consequences for arc magma generation. Earth and Planetary Science Letters, 163, 361-379.

Smith, D.C. (1988) A review of the peculiar mineralogy of the “Norwegian coesite-eclogite province”, with crystal-chemical, petrological, geochemical and geodynamical notes and an extensive bibliography. In Eclogites and Eclogite-Facies Rocks (Smith, D.C. Ed.), pp 546, Elsevier, Amsterdam, 1-206.

Taguchi, T., Enami, M. and Kouketsu, Y. (2016) Prograde evolution of Sulu UHP metamorphic rock in Yangzhuang, Junan region, deduced by combined Raman and petrological studies. Journal of Metamorphic Geology, 34, 683-696.

Tang, J., Zheng, Y.-F., Wu, Y.-B, Gong, B., Zha, X. and Liu, X. (2008) Zircon U-Pb age and geochemical constraints on the tectonic affinity of the Jiaodong terrane in the Sulu orogen, China. Precambrian Research, 161, 389-418.

Tatsumi, Y. (1989) Migration of fluid phases and genesis of basaltic magmas in subduction zones. Journal of Geophysical Research, 94, 4697-4707.

Ulmer, P. (2001) Partial melting in the mantle wedge – the role of H2O in the genesis of mantle-derived ‘arc-related’ magmas. Physics of the Earth and Planetary Interiors, 127, 215-232.

Vitala Brovarone, A. and Agard, P. (2013) True metamorphic iso-grads or tectonically sliced metamorphic sequence? New high-spatial resolution petrological data for the New Caledonia case study. Contributions to Mineralogy and Petrology, 166, 451-469.

Van den Bleeken, G. and Koga, K.T. (2015) Experimentally determined distribution of fluoride and chlorine upon hydrous slab melting, and implications for F-CI cycling through subduction zones. Geochimica et Cosmochimica Acta, 171, 353-373.

Wallis, S.R., Ishiwatari, A., Hirajima, T., Ye, K., Guo, J., Nakamura, D., Kato, T., Zhai, M., Enami, M., Cong, B. and Banno, S. (1997) Occurrence and field relationships of ultrahigh-pressure metagranitoid and coesite eclogite in the Su-Lu terrane, eastern China. Journal of Geological Society, London, 153, 45-54.

Whitney, D.L. and Evans, B.W. (2010) Abbreviations for names of rock-forming minerals. American Mineralogist, 95, 185-187.

Ye, K., Yao, Y., Katayama, I., Cong, B., Wang, Q. and Maruyama, S. (2000) Large areal extent of ultrahigh-pressure rocks from the southern Sulu region, eastern China. Journal of Metamorphic Geology, 13, 659-675.