Eitaro SATO*, Takao HIRAJIMA*, Kenichiro KAMIMURA** and Yoshikazu FUJIMOTO***

*Department of Geology and Mineralogy, Graduate School of Science, Kyoto University, Kitashirakawa Oiwakecho, Sakyo-ku, Kyoto 606-8502, Japan

**Krosaki Harima Co., 1-1 higashihama-cho, Yahatanishi-ku, Kitakyushu, Fukuoka 806-8586, Japan

***Nittetsu Mining Co., Ltd, Yusen Building, 3-2 Marunouchi 2-chome, Chiyoda-ku, Tokyo 100-8377, Japan

White mica K–Ar age dating was carried out for metapelites in the Hakoishi sub-unit and a meta-pillow basalt in the Tobiishi sub-unit of the Kurosegawa belt (Kamimura et al., 2012), Yatsushiro district, Kyushu, Japan.

The Hakoishi sub-unit suffered the lawsonite–blueschist facies metamorphism with peak conditions of 0.7–0.9 GPa and <300 °C. Phengite separate with K content of 5.323 wt% from OD 28 gives 299.0 ± 6.1 Ma and those with K content of 1.222 wt% and 2.142 from OD 113 give 280.2 ± 5.8 Ma and 245.3 ± 5.1 Ma, respectively.

The Tobiishi sub-unit suffered the prehnite–pumpellyite facies metamorphism with the peak conditions round <0.4 GPa and <350 °C. White mica named as alumino-celadonite with a composition range of Si from 3.79 to 3.94 and Fe3+ from 0.30 to 0.48 for Oxygen = 11 basis is developed in amygdules of meta-pillow basalts, closely associated with pumpellyite and chlorite. Alumino-celadonite separates obtained from pillow basalt, SOD01-B, vary K content from 5.601 to 2.343 wt% and give K–Ar ages from 181 to 173 Ma. The peak P–T conditions of the all studied samples were much lower than the closure temperature of Ar in white micas, so these data can be interpreted as the peak metamorphic timing of each sub-unit, i.e., the existence of both late Paleozoic (299–245 Ma) lawsonite blueschist facies metamorphic rocks and Mesozoic (181–173 Ma) prehnite-pumpellyite facies metamorphic rocks are newly identified from the Kurosegawa belt in Kyushu, although Mesozoic epidote–blueschist facies metamorphic rocks have been reported in the relevant area. Obtained metamorphic ages and grade from the Hakoishi and Tobiishi sub-units are almost similar with those of the Osayama area in the Renge belt (Tsujimori and Itaya, 1999) and of the Katsuyama area in the Suo belt of the Inner Zone of Southwest Japan, respectively (Hashimoto, 1968; Nishimura, 1998). These data support the Inner Zone origin of the Kurosegawa belt proposed by Isozaki and Itaya (1991).

Keywords: K–Ar age, Phengite, Alumino-celadonite, Kurosegawa belt, Blueschist

INTRODUCTION

The Kurosegawa belt is distributed as a narrow belt within the northern Chichibu belt composed of the Jurassic accretionary complex in the Outer Zone of Southwest Japan (e.g., Ichikawa et al., 1956; Isozaki and Itaya, 1991). The Kurosegawa belt is mainly constituted of serpentinites, 400 Ma granite and high-temperature metamorphic rocks, high-pressure (HP)/low-temperature (LT) metamorphic rocks with 450–180 Ma age ranges, non-metamorphic Silurian–Devonian sedimentary rocks and Permian accretionary complex (e.g., Isozaki and Itaya, 1991, Isozaki et al., 1992). This belt is characterized by the chaotic occurrence of various rock types with variety of origins in a limited area, i.e., non-metamorphic or metamorphic rocks, lower-geothermal (subduction zone) or higher-geothermal (continental/island arc crust) settings, the sedimentation age contrasts from Silurian to Jurassic, and etc. To explain the origin of such complicated geology, several models have been proposed; 1) either a fragment of large strike–slip fault system accompanied with serpentinite (Taira et al., 1983; Maruyama et al., 1984) or 2) extended nappe originated from the Inner Zone of
Southwestern Japan (Isozaki and Itaya, 1991).

Isozaki and Itaya (1991) and Isozaki et al. (1992) pointed out the similarity of the white mica K–Ar ages of HP/LT metamorphic rocks among the Inner Zone (430–280, 230–150 Ma) and the Kurosegawa belt (445–300, 240–180 Ma), and they proposed that the Kurosegawa belt was a tectonic outlier of the Inner Zone. Nishimura (1998) divided the so-called Sangun metamorphic rocks in the Inner Zone of Southwest Japan into two belts; Renge and Suo belts using the age data reported. The former is characterized by 330–280 Ma ages and HP/LT types metamorphic rocks such as the blueschist (BS) to epidote-amphibolite (EA) facies metamorphic rocks. The Suo belt is characterized by 230–160 Ma age range and HP/LT metamorphic rocks related from pumpellyite–actinolite (PA) through BS to EA facies.

Recent U–Pb ages of detrital zircons extracted from metasediments and quartzites closely associated with HP/LT type metamorphic rocks in the Kurosegawa belt show four age distribution concentrated at 450–500 Ma and 600 Ma, a scattering from 800 to 1800 Ma, and older than 2000 Ma (Yoshimoto et al., 2013). Such age distribution of detrital zircons is identical with those from Permian–Triassic collision–related metamorphic rocks of the Ogcheon metamorphic belt and Gyeonggi massif in Korea, Japan, and from Devonian sedimentary rocks in the South China Craton. Therefore, they proposed that the origin of the Kurosegawa belt is closely correlated with the development history in the Eastern margin of Eurasia continent.

In Kyushu, the Kurosegawa belt is exposed to the south of the Usuki-Yatsushiro Tectonic Line (e.g., Ichikawa et al., 1956) as a narrow belt with the width up to 20 km (Fig. 1a). In this belt, HP/LT metamorphic rocks are exposed as blocks in serpentinite mélange of Hakoishi, Tobishi, Fukami, Haki, Gokanosho and Fukami areas from north to south (Fig. 1a; Matsumoto and Kanmera, 1964; Saito et al., 2005). Nishizono (1996) and Saito et al. (2005) have reported K–Ar ages of HP/LT metamorphic rocks in the Hakoishi and Gokanosho areas, respectively, and their results are concentrated between 187 Ma and 144 Ma for metasediments, showing a close age–relationship with the Suo belt. However, the K–Ar age of metamorphic rocks related with the Renge belt has not been reported yet from the Kurosegawa belt in Kyushu.

The preliminary K–Ar age dating was carried out for two white mica separates from lawsonite–blueschist (LBS) facies metapelites in the Hakoishi area and celadonite separate from a prehnite–pumpellyite (PrP) facies meta–pillow basalt in the Tobishi area (Fig. 1b). We firstly obtained the K–Ar age related with the Renge belt from the LBS facies metamorphic rocks in the Kurosegawa belt of Kyushu. In this paper, we will report the detailed results of K–Ar dating and petrography especially focused on white micas and will discuss their significance on the origin of the Kurosegawa belt.

GEOLOGIC AND METAMORPHIC BACKGROUND OF THE HAKOISHI AND TOBIISHI SUB-UNITs

The Kurosegawa belt in the Yatsushiro district mainly consists of serpentinite mélangé including HP/LT type metamorphic rocks, Carboniferous limestone, Paleozoic–Mesozoic sedimentary rocks and Permian–Jurassic accretionary complex along with minor amount of granite block and pyroxenite (Kanmera, 1952; Ichikawa et al., 1956; Saito et al., 2005) (Fig. 1a).

In the Hakoishi and Tobishi areas of the Yatsushiro district (Fig. 1b), some geological/metamorphic interpretations have been proposed. Kanmera (1952) reported the basic rock dominant layer (Tobishi Formation) continues more than 20 km in east–west trend accompanied with the sedimentary rocks dominant layer (Shimotake Formation). However, Saito et al. (2005) proposed that the western half of the Tobishi Formation occupied the major part of Hakoishi serpentinite mélangé, and the eastern half of the Tobishi Formation and the Shimotake Formation are classified to the Otao unit.

Ueta (1961) firstly reported the occurrence of LBS from the Hakoishi area. He divided the Tobishi Formation and Shimotake Formation of Kanmera (1952) into two metamorphic zones based on the mineral assemblages of metabasites; zone 1 characterized by the occurrence of LBS corresponding with the western half of the Tobishi Formation, and zone 2 characterized by the greenschist facies mineral assemblage corresponding with the eastern half of the Tobishi Formation and Shimotake Formation. He proposed that the metamorphic grade increased from the zone 1 to zone 2 and then concluded that these metamorphic rocks could be the western extension of the Sanbagawa belt based on the limited geological and petrological knowledge at that time.

Since his pioneering work, petrological studies were not carried out in the Hakoishi and Tobishi area. Recently, we confirmed that the LBS reported by Ueta (1961) is extremely fresh and almost free from hydration reactions suffered during the exhumation stage, which are common in the Sanbagawa metamorphic rocks (e.g., Uno et al., 2014). Kamimura et al. (2012) confirmed that LBS is predominant in the zone 1 of Ueta (1961), but prehnite and/or pumpellylite occur in metabasites in zone 2 of Ueta (1961) except for the Shimotake Formation, and then
they proposed a new metamorphic zonal mapping, such as LBS zone mainly occurred in the Hakoishi sub-unit and PrP facies unit mainly developed in the Tobiishi sub-unit, corresponding with the eastern half of the Tobiishi Formation of Kanmera (1952). As metabasites of LBS and PrP facies rocks are generally lacking in the Sanbagawa belt, they considered that these metabasites belong to the Kurosegawa belt.

Saito et al. (2005) named the western half of the Tobiishi Formation of Kanmera (1952) as Hakoishi serpentinite unit. They considered that LBS body in the Hakoishi serpentinite unit as tectonic blocks in serpentinite. However, our systematic sampling in the relevant area identified that serpentinite is mainly distributed at the southern edge of LBS facies metamorphic body. Therefore, we consider that the LBS facies metamorphic rocks in Hakoishi area (Hakoishi sub-unit) form a coherent block up to 10 km long, based on geological aspects mentioned later.

According to our survey, the Hakoishi sub-unit is mainly composed of metabasites, and metachert with minor amount of metagabbro and metapelite. All of these rocks record the LBS facies metamorphism, represented by the occurrence of lawsonite, sodic amphibole, sodic pyroxene and/or albite. The original igneous textures of hyaloclastite and pillow breccia are well preserved, although a schistose structure is developed in some metabasites in the western part of the Hakoishi sub-unit (Fujimoto et al., 2010). The chert shows various colors in the field such as white, red to pinkish, pale–green, brown and black depending on metamorphic minerals (Ibuki et al., 2008, 2010), and commonly shows banding structures with a few cm thick and complicated folding structures, probably representing a slump structure. Metagabbro containing jadeite but without quartz is also reported by Saito and Miyazaki (2006). Their peak metamorphic con-
ditions are estimated as 0.7–0.9 GPa and <350 °C based on the mineral assemblages (Liou et al., 1985; Banno, 1998).

The Tobiishi sub-unit is mainly composed of basic rocks and Yayama-dake limestone with minor amount of amphibolite, acidic tuff, chert and sedimentary rocks. Basic rocks commonly preserve the original igneous structures, such as pillow–structure, pillow breccia and hyaloclastite, and the penetrative foliation cannot be visible in the field. The metamorphic minerals are mainly developed in veins and amygdules of basic rocks, and they are mainly pumpellyite, prehnite, chlorite, titanite and greenish–mica with minor amount of epidote and calcite. Greenish–mica is identified only in amygdules in pillow lava. These minerals are identified with PrP facies metamorphism, suggesting the peak metamorphic temperature is less than 300 °C (Liou et al., 1985; Banno, 1998).

PETROGRAPHY

Hakoishi sub-unit

Through our survey, metapelites were found from a few outcrops in the Hakoishi sub-unit. We selected OD28 and OD113 collected from the eastern and western edges of the Hakoishi sub-unit for geochronological studies. These rocks are rich in white mica with close association with lawsonite, sodic amphibole and sodic pyroxene, suggesting a coeval origin with other high-pressure minerals of the LBS facies metamorphism (Figs. 1a and 2a–2d).

Sample OD28 is composed of intercalations of quartz-rich, sodic pyroxene–white mica–rich and sodic amphibole–white mica–rich layers with several millimeters width. Quartz grains in the quartz–rich layer generally show a wavy extinction. This rock is mainly composed of white mica, sodic amphibole, sodic pyroxene (mainly augite to jadeite with Jd40–61Ac30–40Di4–9), albite, quartz and titanite along with minor amounts of chlorite, apatite and zircon. Sodic pyroxene mainly occurs as a fine-grained tabular crystal, up to 30 µm in width, with colorless to bright yellowish orange color under the plane polarized light. Sodic amphibole occurs as needle to columnar shaped crystals with ca. 10–100 µm in the long axis characterized by the pleochroisms of colorless, pale purple and pale blue. White mica also occurs as very fine–grained minerals ca. 10–100 µm in the long axis with close association with other high–pressure minerals and has a composition range of Si = 3.55–3.64 for O (Oxygen) = 11 basis (Fig. 3a). Hereafter, white mica with Si content from 3.25 to 3.75 will be tentatively called as phengite. The modal amount of phengite is the highest among the main constituents, except for quartz. Most of phengite grains are free from paragonite component (Table 1) and are plotted around the tie line between ideal muscovite and ideal aluminoceladonite in Si–Al and Si–(Fetotal + Mg) diagrams (Figs. 3a and 3b). This means that the phengite in OD28 is almost free from Fe3+ content and the chemical variation was caused by the tschermak substitution Mg13Si4Al4 = Al13Al4P. Some phengite grains show a pale brownish color at the margin, and they are plotted slightly apart from the tie line between ideal muscovite and ideal aluminoceladonite (Fig. 3a).

The sample OD113 is also composed of an intercalation of three different colored layers with several millimeters width; white, grayish yellow and dull blue layers. The white color layer is mainly composed of quartz and carbonate mineral and quartz grains in this layer are mainly polygonal with ~10 µm in diameter. The grayish yellow layer is mainly composed of fine–grained sodic amphibole, lawsonite, and white mica with minor amount Cr–pumpellyite, zircon, chromite and calcite. Sodic amphibole and lawsonite occur generally as needle or tabular shaped in the ~10–150 µm and ~30–100 µm with in the long dimension, respectively. The dull blue layer is characterized by higher modal amount of sodic amphibole rather than those in the grayish yellow layer. The occurrence of relic chromite and chromian–pumpellyite and the abovementioned millimeters scale layering structure in the sample OD113 suggest that its protolith should be a mixture of ultramafic/mafic materials and siliceous ooze in a past ocean floor or subduction zone.

In the sample OD113, most white mica occurs in the white mica–rich domain of the grayish yellow layer (Fig. 2d). White mica grains occupied in the central part of white mica–rich domain are colorless, but those located at the margin of white mica–rich domain show a pale brownish color. The composition of colorless white mica grains are plotted around the tie line between ideal muscovite and ideal aluminoceladonite in Si–Al and Si–(Fetotal + Mg) diagrams (Fig. 3) with Si range from 3.63 to 3.72. Si content of white mica with a pale brownish color ranges from 3.35 to 3.62. However, their compositions are plotted apart from the abovementioned tie line and are aligned along a tie line between phengite solid solution and chlorite in the same sample (Fig. 3b). These observations suggest that the pale brownish color of phengite reflects the submicroscopic chloritization of phengite during the later stage, although the backscatter observation under the electron microscope cannot detect the chloritization effect on phengite.

Tobiishi sub-unit

We selected meta–pillow basalt, SOD01–B, in the Tobi-
Figure 2. (a) A photomicrograph of pelitic schist of OD28 in the Hakoishi sub-unit. Phengite (Phn) is one of main foliation forming mineral. Qtz, quartz; Naam, sodic amphibole. Crossed nicols. (b) A backscattered image of Phn, sodic pyroxene (Napx) and Naam in OD28. (c) A photomicrograph of pelitic schist of OD113 in the Hakoishi sub-unit. Phengite coexists with Lws, Chl and Qtz. Crossed nicols. (d) A backscattered image of OD113 shows close association with Phn, Lws and Naam. (e) A photomicrograph of amygdules. Each amygdule is commonly occupied by metamorphic minerals such as alumino-celadonite (Cel), Chl and Qtz in a meta-pillow basalt of SOD01-B. (f) A backscattered image of amygdule in which metamorphic minerals formed a layered structure, i.e., alumino-celadonite core, chlorite mantle and pumpellyite (Pmp) rim. The matrix is characterized by the dendritic texture of clinopyroxene and plagioclase, suggesting the rapid crystal growth. Color version is available online from http://japanlinkcenter.org/DN/JST/JSTAGE/jmps/140820.
ishi sub-unit for dating [Fig. 2b of Kamimura et al. (2012)], because this rock has many amygdules, ~ 2.0–0.2 mm in diameter, including greenish-mica (Figs. 1b and 2e-2f). Under the microscope, dendritic igneous clinopyroxene and plagioclase occupy the matrix of the rock. The metamorphic minerals of chlorite, greenish-mica, quartz and albite are developed mainly in amygdules and veins, in which following assemblages are identified; greenish-mica + chlorite + quartz, greenish-mica + chlorite + quartz + pumpellyite, greenish-mica + calcite + chlorite, chlorite + quartz, chlorite only.

About half of amygdules form layered structure composed of chlorite rim, chlorite + greenish-mica mantle and greenish-mica core (Fig. 2e). Some layered amygdules consist of pumpellyite rim, chlorite mantle and greenish-mica core (Fig. 2f). Small amygdules less than 100 µm were almost composed of chlorite or chlorite + quartz (Fig. 2d). Rarely, calcite core was observed. Pumpellyite occurs both in amygdules and veins as columnar shaped grain or aggregate with polysynthetic twining structure with ~ 30–10 µm length. Greenish-mica generally occurs as very fine-grained minerals ~ 5–10 µm in amygdules. Almost greenish-mica are associated with pumpellyite, chlorite and quartz and have following chemical composition range; Si = 3.79–3.94, Fe³⁺ = 0.30–0.48 for O = 11 (Table 1), so named as aluminocaladonite (Rieder et al., 1999). Compositions of greenish-mica are plotted apart from the tie-line of ideal muscovite and ideal aluminocaladonite, but they are aligned almost parallel to the tie-line but to not toward chlorite (Fig. 3). Therefore, these chemical characters reflect the incorporation of significant amount of Fe³⁺ content in the

Figure 3. (a) Si-Al diagram of white micas (Oxygen = 11). Cross and diamond symbols represent compositions of white micas of OD28 and OD113, respectively, in the Hakoishi sub-unit. Open circle indicates compositions of greenish-micas of SOD01-B in the Tobiishi sub-unit. (b) Si - (Fe* + Mg) diagram of white micas and chlorite (Oxygen = 11). Symbols are same as in Figure 3a, except for solid circle showing composition of chlorite in OD113. Fe* = Fe total.
studied alumino-celadonite (Table 1). As the natural alumino-celadonite often coexists with pumpellyite in amygdules, the alumino-celadonite can be regarded as the product of the PrP facies metamorphism.

### K-AR AGES OF WHITE MICAS

The rock samples of OD28 and OD113 were crushed and sieved. The sieved fractions of 50–100 µm in OD28 and of 50–75µm in OD113 were firstly prepared based on the grain size of phengite in each rock. Phengite is concentrated from these sieved fractions using isodynamic separator, heavy liquid and tapping technique following Yagi (2006) and Yagi and Itaya (2011). For OD113, the phengite separate of 0.5–2.0 µm is further prepared using a centrifuge, as the first phengite separate of OD113 shows very low K content (ref., Table 2). For SOD01-B, the rock sample was crushed using the High Voltage Pulse Selective Crushing Equipment (selFrag Lab) installed in JAMSTEC (e.g., Wang et al., 2011) and then 30 g of amygdules were collected by the hand picking. Alumino-celadonite was concentrated from amygdules in Hiruzen Institute for Geology and Chronology Co. Ltd. The amygdule separates were crushed by an iron mortar and then a fraction of 2–50 µm was prepared using a sieve and a centrifuge. The fraction was treated with 2.5N HCl solution for 30 minutes on hot plate (70 °C) and washed with pure water repeatedly. The fraction with water in a breaker makes a grading in color from green in bottom to greenish yellow in upper. Three fractions, green in the bottom part, greenish yellow in the top and a mixture in the middle were separated for dating.

Analysis of potassium and argon of white mica separates, and calculations of ages and errors were carried out following the methods described by Nagao et al. (1984) and Itaya et al. (1991). Potassium was analyzed by flame photometry using a 2000 ppm Cs buffer with an analytical error within 2% at a 2-sigma confidence level. Argon was analyzed on a 15 cm radius sector type mass spectrometer with a single collector system using the isotopic dilution method and 38Ar spike. Multiple runs of the standard (JG–1 biotite, 91 Ma) indicate that the error of argon analysis is about 1% at a 2-sigma confidence level (Itaya et al., 1991). The decay constants of $^{40}$K to

### Table 1. Representative chemical compositions of phengite and alumino-celadonite

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Sample No.</th>
<th>OD28</th>
<th>OD113</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe$^{2+}$</td>
<td>0.10</td>
<td>0.10</td>
<td>0.12</td>
</tr>
<tr>
<td>Mn</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>Mg</td>
<td>0.40</td>
<td>0.43</td>
<td>0.41</td>
</tr>
<tr>
<td>Ca</td>
<td>0.00</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Na</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>K</td>
<td>0.04</td>
<td>0.03</td>
<td>0.95</td>
</tr>
</tbody>
</table>
$^{40}$Ar and $^{40}$Ca, and $^{40}$K content in potassium used in the age calculations are $0.581 \times 10^{-10}$/year, $4.962 \times 10^{-10}$/year and 0.0001167, respectively (Steiger and Jäger, 1977).

The K-Ar dating results are shown in Table 2. In the Hakoishi sub-unit, 299.0 ± 6.1 Ma was obtained for OD28, and 280.2 ± 2.7 and 245.3 ± 3.9 Ma for OD113. The phengite separate has moderate K content for OD28 (5.323 wt%) but two phengite separates for OD113 have as low K content as 1.222 and 2.142 wt%, which should be caused by the abovementioned chloritization of phengite and the presence of fine-grained high-pressure minerals such as sodic amphibole (Fig. 2d).

Grain sizes of dated aluminoceladonite for SOD01-B in the Tobiishi sub-unit are less than 50 µm and K contents of aluminoceladonite separates range from 5.601 to 2.343 wt% (Table 2). In spite of the variable K content in separates, obtained K-Ar ages are concentrated around 181–173 Ma (Table 1).}

**DISCUSSION**

The phengite K-Ar ages of 299–245 Ma obtained from the LBS facies metamorphic rocks in the Hakoishi sub-unit by this study obviously differ from Mesozoic white mica ages from PA and epidote-blueschist (EBS) facies metamorphic rocks previously reported in the Yatsushiro district of the Kurosegawa belt (Figs. 1a and 4), i.e., 149–187 Ma from the Haki metamorphic rocks (Nishizono, 1996) and 182–144 from the Gokanosho metamorphic
rocks (Saito et al., 2005). Our new result and the previous K–Ar ages were all obtained from white mica separates extracted from the metamorphic rocks of which peak P–T conditions are significantly lower than the closure temperature of the K–Ar white mica system (~ 500 °C; e.g., Itaya et al., 2011), so there is a fear that the obtained ages could be inherited from older detrital/protolith white micas.

Our optical microscopic and EPMA analyses for two metapelite samples revealed that phengites in both samples occur as one of main foliation forming minerals and the non–chloritized phengites are generally homogeneous in each grain with Si contents ranging from 3.55 to 3.64 (O = 11) for OD28 and from 3.63 to 3.72 (O = 11) for OD113 (Fig. 3a). There is a tendency that the alumino–celadonite substitution [(Mg,Fe)Si = AlVIAlIV] in white mica is enhanced under higher-pressure and lower-temperature conditions of the blueschist facies metamorphism, e.g., Si content of white mica in metapelites of the chlorite zone of the Sanbagawa belt, PA facies equivalent, varies from 3.3 to 3.5 and this value decreases to Si = 3.2–3.4 in the biotite zone, equivalent to the EA facies (e.g., Hirajima et al., 1992). Si content of white mica varies from 3.4–3.6 in metapelites of LBS facies in the Renge belt (Tsujimori and Itaya, 1999). Compared with these reported value and the mode of occurrence, the phengite in our studied samples were certainly formed under the blueschist facies metamorphism, and that the obtained K–Ar ages are representative of the peak metamorphic age of the Hakoishi sub-unit.

Previous studies reported a tendency that the white mica–separate with low K content sometimes gives younger ages due to the effect of impurities in the separation (e.g., Tsujimori and Itaya, 1999: Itaya et al., 2011). Natural white mica in high-pressure metamorphic rocks contains normally 10–11 wt% of K2O. Three phengite separates from OD113 (K2O = 1.47 wt% and 2.58 wt%) and OD 28 (K2O = 6.41 wt%) are having significantly lower than the expected K2O value of white mica. The phengite separate with the highest concentration of K2O shows the oldest age. Therefore, we consider phengite K–Ar age of 299 Ma can be more reliable rather than 280 and 245 Ma of OD113, although slightly K2O poor separate gives an older age (280 Ma) than the richer separate (245 Ma).

Metamorphic alumino–celadonite commonly occurs in amygdules and veins in basaltic or andesitic rocks suffered from the zeolite to PrP facies metamorphism. A limited number of K–Ar dating for alumino–celadonite has been published by previous workers. Oliveros et al. (2008) applied the K–Ar method for metamorphic celadonite and U–Pb method for the titanite in metabasites suffered the PrP facies metamorphism in the Andes of central Chile. Celadonite is identified from amygdules along with titanite and quartz [see Figs. 2a and 2b of Oliveros et al. (2008)]. The celadonite K–Ar ages (109–82 Ma) are almost similar with the U–Pb age of titanite (106–84 Ma). Therefore, they concluded that celadonite and titanite should be reliable tools of dating and that both ages showed metamorphic timing of PrP facies metamorphism. Therefore, we consider that our alumino–celadonite K–Ar ages (173–181 Ma for SOD01–B) as PrP facies metamorphic timing, which are correlatable with the Mesozoic the Haki and Gokanosho metamorphic rocks in the relevant area (Nishizono, 1996; Saito et al., 2005).

Abovementioned our data suggest that two types of HP/LT type metamorphic rocks, late Paleozoic LBS facies metamorphic rocks and Mesozoic metamorphic rocks from PrP facies through PA facies to EBS-facies, are exposed in the Yatsushiro district of the Kurosegawa belt.

Saito et al. (2005) pointed out a possibility that the Haki and Gokanosho metamorphic rocks could be correlated with each other based on their resemblance of metamorphic grade (PA to EBS facies) and metamorphic age (~ 190–140 Ma), although they are distributed in separate areas (Fig. 1a). Furthermore, Mesozoic white mica K–Ar ages of 217–206 Ma were newly reported from metasediments associated with PA to EBS facies metabasites from the Kurosegawa belt in eastern Kyushu (Miyazaki et al., 2014). In spite of the similarity of K–Ar ages among the Tobiishi sub–unit and Mesozoic metamorphic rocks from other areas of the Kurosegawa belt in Kyushu, metamorphic grade of PrP facies in the Tobiishi sub–unit is lower than those of abovementioned metamorphic rocks. Therefore, the Tobiishi sub–unit has some contrasting characteristics when compared with the Mesozoic high pressure type metamorphic rocks of the Kurosegawa belt in Kyushu. This would probably mean that the Tobiishi sub–unit may represent the shallower part of subducting plate during the middle Mesozoic time and other Mesozoic high pressure type metamorphic rocks may occupy the deeper part of the same subduction system.

Obtained white mica K–Ar ages and metamorphic facies of the Hakoishi and Tobiishi sub–unit correspond with those of the Renge and Suo belt, respectively (Nishimura, 1998). To consider the genetic relationship between HP/LT metamorphic rocks in the Inner Zone of Southwest Japan, all of available K–Ar white mica ages are plotted on a geological map (Fig. 5 and Table 3).

Phengite K–Ar ages of 273–327 Ma were reported from metapelites and LBS found from the Osayama serpentinite melange of the Renge belt located in Chugoku Mountains (Tsujimori and Itaya, 1999). The age range
and the metamorphic facies of the studied LBS are comparable with those of the LBS in Osayama serpentinite mélangé of the Renge belt, although LBS and associated rocks found from the Ino Formation of the Kurosegawa belt in Shikoku gave an intermediate K–Ar white mica age between the Suo and Renge metamorphism ranging from 240 to 208, 394 to 317 Ma (Maruyama et al., 1978; Ueda et al., 1980).

PrP facies metabasite has been reported from the Suo belt. According to Hashimoto (1968), Katsuyama area in Chugoku Mountains can be divided into three metamorphic zones, i.e., Zone I of PrP facies metamorphism characterized by the occurrence of prehnite, pumpellyite, chlorite and white mica with minor amount of epidote, actinolite in metabasites, although distinct metamorphic minerals were not identified from associated sedimentary rocks nor limestone with fusulinid fossils, Zone II of PA or EBS facies metamorphism characterized by the occurrence of sodic amphibole, epidote, actinolite, pumpellyite and chlorite in basic–schists, and Zone III of greenschist facies metamorphism characterized by the occurrence of actinolite, epidote and chlorite in basic–schists. Subsequently, Shibata and Nishimura (1989) reported a white mica K–Ar age of 191 ± 6 Ma from a metabasite containing pumpellyite and actinolite, which was collected from Zone II of Hashimoto (1968). Lithotype, metamorphic age and its grade of the Katsuyama area are almost identical with those of the Tobiishi sub-unit and Haki/Gokanosho metamorphic rocks.

According to the geotectonic cross-section proposed by Saito et al. (2005) (their Fig. 9.1), the Hakoishi sub-unit and the accompanying serpentinite are underlain by the Tobiishi sub-unit and the accompanying Yayama–dake limestone body (Fig. 1b). In Inner Zone of Southwest Japan, older blueschist of the Renge belt commonly occupies a higher structural level as a klippe than those of younger Suo metamorphic rocks (Nishimura, 1998; Tsujimori and Itaya, 1999), which is similar to the geo–structural relationship between the Renge and Suo belts. These data also support the Kurosegawa klippe model of Isozaki and Itaya (1991).

Our geochronological and petrological studies in the Kurosegawa belt in the Yatsushiro district revealed that the Hakoishi sub–unit can be formed by the Carboniferous to Permian subduction of an oceanic plate, and the Tobiishi sub–unit and the Haki/Gokanosho metamorphic rocks can be formed by the Jurassic subduction. Osanai et al. (2014) reported 492.4 ± 3.6Ma U–Pb concordia age of zircons extracted from a glaucophane bearing metabasalt from the Hakoishi sub–unit and they interpreted that this age represented the crystallization timing of the mid–oceanic–ridge type basaltic magma in the oceanic crust. If so, there is a huge time gap of about 200 Ma between the oceanic crust formation and the subsequent subduction which caused LBS facies metamorphism. This would suggest the fairly old and cold plate subduction could make the very lower geothermal gradient to form LBS such as the Hakoishi sub–unit.
Table 3. Detailed K-Ar age and rock type of the Renge and Suo belt in Southwest Japan

<table>
<thead>
<tr>
<th>No.</th>
<th>Tectonic belt</th>
<th>Loc. Name</th>
<th>Age (Ma)</th>
<th>Facies</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Kurosegawa (K)</td>
<td>Hakoishi</td>
<td>299-245</td>
<td>LBS</td>
<td>this study</td>
</tr>
<tr>
<td>2</td>
<td>Kurosegawa (K)</td>
<td>Tobishi</td>
<td>181-173</td>
<td>PrP</td>
<td>this study</td>
</tr>
<tr>
<td>3</td>
<td>Kurosegawa (K)</td>
<td>Gokanosho</td>
<td>182-144</td>
<td>PA-EBS</td>
<td>Saito et al. (2005)</td>
</tr>
<tr>
<td>4</td>
<td>Kurosegawa (K)</td>
<td>Haki</td>
<td>187-149</td>
<td>PA-EBS</td>
<td>Nishizono (1996)</td>
</tr>
<tr>
<td>5</td>
<td>Kurosegawa (K)</td>
<td>Mie-city</td>
<td>217±5</td>
<td>PA-EBS</td>
<td>Miyazaki et al. (2014)</td>
</tr>
<tr>
<td>6</td>
<td>Kurosegawa (K)</td>
<td>Usuki</td>
<td>206±5</td>
<td>PA-EBS</td>
<td>Miyazaki et al. (2014)</td>
</tr>
<tr>
<td>7</td>
<td>Renge (K)</td>
<td>Kiyama</td>
<td>337-290</td>
<td>EBS-EA</td>
<td>Ueda and Onuki (1968); Kabashima et al. (1995)</td>
</tr>
<tr>
<td>8</td>
<td>Renge (K)</td>
<td>Wakamiya</td>
<td>272-256</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>9</td>
<td>Suo (K)</td>
<td>Manotani</td>
<td>214-168</td>
<td>PA</td>
<td>Okamoto et al. (1989); Nagakawa et al. (1997)</td>
</tr>
<tr>
<td>10</td>
<td>Suo (K)</td>
<td>Yamaga</td>
<td>193±6</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>11</td>
<td>Suo (K)</td>
<td>Kurume</td>
<td>184-159</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>12</td>
<td>Suo (K)</td>
<td>Yame</td>
<td>218-211</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>13</td>
<td>Kurosegawa (H)</td>
<td>Toba</td>
<td>209-192</td>
<td>PA</td>
<td>Isozaki and Itaya (1992)</td>
</tr>
<tr>
<td>15</td>
<td>Renge (H)</td>
<td>Osayama</td>
<td>326-273</td>
<td>LBS,EBS</td>
<td>Tsujimori and Itaya (1999)</td>
</tr>
<tr>
<td>16</td>
<td>Suo (H)</td>
<td>Yamaguchii</td>
<td>206±7</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>17</td>
<td>Suo (H)</td>
<td>Nishiki-cho</td>
<td>228±7</td>
<td>PA-EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>18</td>
<td>Suo (H)</td>
<td>Ikura</td>
<td>190-170</td>
<td>PA-EBS</td>
<td>Nishimura (1998)</td>
</tr>
<tr>
<td>19</td>
<td>Suo (H)</td>
<td>Gotsu</td>
<td>195-191</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>20</td>
<td>Suo (H)</td>
<td>Tsukita</td>
<td>165±5</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>21</td>
<td>Suo (H)</td>
<td>Katsuyama</td>
<td>191±6</td>
<td>PrP-PA</td>
<td>Shibata and Nishimura (1989); Hashimoto (1968)</td>
</tr>
<tr>
<td>22</td>
<td>Suo (H)</td>
<td>Asahi-cho</td>
<td>181-120</td>
<td>EBS</td>
<td>Shibata and Nishimura (1989)</td>
</tr>
<tr>
<td>24</td>
<td>Kurosegawa (S)</td>
<td>Shirokawa</td>
<td>277-265</td>
<td>PA</td>
<td>Isozaki and Itaya (1992)</td>
</tr>
<tr>
<td>25</td>
<td>Kurosegawa (S)</td>
<td>Nakatsuyama</td>
<td>233-179</td>
<td>PA</td>
<td>Isozaki and Itaya (1991); Aiba (1982)</td>
</tr>
<tr>
<td>26</td>
<td>Kurosegawa (S)</td>
<td>Agekura</td>
<td>228-185</td>
<td>Unknown</td>
<td>Isozaki and Itaya (1990)</td>
</tr>
<tr>
<td>27</td>
<td>Kurosegawa (S)</td>
<td>Kamikatsu</td>
<td>225-194</td>
<td>PA-EBS</td>
<td>Suzuki et al. (1990)</td>
</tr>
<tr>
<td>28</td>
<td>Kurosegawa (S)</td>
<td>Ino</td>
<td>240-204,394-317</td>
<td>LBS</td>
<td>Maruyama et al. (1978); Ueda et al. (1980)</td>
</tr>
<tr>
<td>29</td>
<td>Kurosegawa (S)</td>
<td>Sawadani</td>
<td>215-205</td>
<td>PA</td>
<td>Isozaki and Itaya (1992)</td>
</tr>
<tr>
<td>30</td>
<td>Kurosegawa (S)</td>
<td>Shimizu-cho</td>
<td>217-198</td>
<td>PA</td>
<td>Isozaki and Itaya (1992); Kurimoto (1993)</td>
</tr>
</tbody>
</table>

Numbers are linked to the geological map of Figure 5.
K, Kurosegawa; S, Shikoku Island; H, Honshu; PrP, prehnite-pumpellyte facies; PA, pumpellyite-actinolite facies; LBS, lawsonite-blueschist facies; EBS, epidote-blueschist facies; EA, epidote-amphibolite facies; S.F., Shitani Formation; H.F., Hattou Formation.

acknowledgments

We would like to express our gratitude to Dr. Kenichiro Tani for providing the opportunity to use the High Voltage Pulse Selective Crushing Equipment at JAMSTEC. We also would like to thank M. Saito, K. Miyazaki, T. Nakajima for their fruitful discussions. This paper benefited from constructive reviews by T. Itaya, Y. Isozaki, and T. Tsujimori and the editorial handling by M. Satish Kumar. This work was supported by Grants-in-Aid for Scientific Research No. 22244067 and No. 25257208.

Supplementary Material

Color version of Figures 2 is available online from http://japanlinkcenter.org/DN/JST/JSTAGE/jmps/140820.

references

Hashimoto, M. (1968) Glaucophanitic metamorphism of the Katsuyama district, Okayama Prefecture, Japan. Journal of the Faciles
Kanmera, K. (1952) The Upper Carboniferous and the Lower Per-
Ar ages of metamorphic and plutonic rocks in the area of 1:200,000 quadrangle geological map of Oita district. Ab-
rocks of the Inner Zone (Untersuchugen über das Chichibu-Terrain in Shikoku-III). The Journal of the Geologi-
Nishizono, Y. (1996) Mesozoic convergent process of the Southern Chichibu Terrane in West Kyushu, Japan, on the basis of Tri-
Nishimura, Y. (1996) Mesozoic convergent process of the Southern Chichibu Terrane in West Kyushu, Japan, on the basis of Tri-
Kurosegawa Belt in west-central Shikoku, Southwest Japan —Kurosegawa Terran as a tectonic outlier of the pre-Jurassic rocks of the Inner Zone—. The Journal of the Geological So-
Itaya, T., Ito, H. and Nagao, K. (1991) Degassing of argon from young geological materials by low temperature stepwise heat-
glaucophane schists in the Kurosegawa tectonic zone near Kochi city, Shikoku. The Journal of the Japanese Asso-
ciation of Mineralogists, Petrologists and Economic Geolo-
gists, 73, 300-310.
Matsumoto, T. and Kammera, K. (1964) Geology of Hinagu dis-
Yatsushiro City, Kumamoto Prefecture, central Kyushu. Bul-
sulty of Science, the University of Tokyo, Section II, 17, 99-
sis of Mn-bearing lawsonite occurring in meta-siliceous rocks in Hakoishi serpentinite mélange of Kurosegawa Belt, Central Kyushu, Japan. Journal of Mineralogical and Petrological Sci-
ences, 105, 340-345.
cal Society of Japan, 62, 82-103 (in Japanese with German abstract).
Isozaki, Y. and Itaya, T. (1999) K-Ar ages of weakly metamor-
Isozaki, Y. and Itaya, T. (1991) Pre-Jurassic klippe in northern Chichibu Belt in west-central Shikoku, Southwest Japan —Kurosegawa Terran as a tectonic outlier of the pre-Jurassic rocks of the Inner Zone—. The Journal of the Geological So-
ciety of Japan, 97, 431-450 (in Japanese with English ab-
strac-
Itaya, T., Ito, H. and Nagao, K. (1991) Degassing of argon from young geological materials by low temperature stepwise heat-
Kabashima, T., Isozaki, Y., Nishimura, Y. and Itaya, T. (1995) Re-
examination on K-Ar ages of the Kiyama high-P/T schists in central Kyushu. The Journal of the Geological Society of Ja-
Kanmera, K. (1952) The Upper Carboniferous and the Lower Perm-
Kurimoto, C. (1993) K-Ar ages of the rocks of Sanbagawa, Kurosegawa and Shimanto Terranes in the northeastern part of Wa-
strac-
Maruyama, S., Ueda, Y. and Banno, S. (1978) 208-240 M.Y. old jadeite-glauconephite schists in the Kurosegawa tectonic zone near Kochi city, Shikoku. The Journal of the Japanese Asso-
ciation of Mineralogists, Petrologists and Economic Geolo-
gists, 73, 300-310.
gists, 36, 905-912.
strac-


Manuscript received August 20, 2014
Manuscript accepted November 1, 2014
Published online December 5, 2014
Manuscript handled by M. Satish-Kumar