Metamorphic evolution of kyanite–staurolite-bearing epidote–amphibolite from the Early Palaeozoic Oeyama belt, SW Japan

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ABSTRACT

Early Palaeozoic kyanite–staurolite-bearing epidote–amphibolites including foliated epidote–amphibolite (FEA), and nonfoliated leucocratic or melanocratic metagabbros (LMG, MMG), occur in the Fuko Pass metacumulate unit (FPM) of the Oeyama belt, SW Japan. Microtextural relationships and mineral chemistry define three metamorphic stages: relict granulite facies metamorphism (M1), high-P (HP) epidote–amphibolite facies metamorphism (M2), and retrogression (M3). M1 is preserved as relict Al-rich diopside (up to 8.5 wt.% Al2O3) and pseudomorphs after spinel and plagioclase in the MMG, suggesting a medium-P granulate facies condition (0.8–1.3 GPa at > 850 °C). An unusually low-variance M2 assemblage, Hbl + Czo + Ky ± St + Pg + Rt ± Ab ± Crn, occurs in the matrix of all rock types. The presence of relic plagioclase inclusions in M2 kyanite associated with clinozoisite indicates a hydration reaction to form the kyanite-bearing M2 assemblage during cooling. The corundum-bearing phase equilibria constrain a qualitative metamorphic P–T condition of 1.1–1.9 GPa at 550–800 °C for M2. The M2 minerals were locally replaced by M3 margarite, paragonite, plagioclase and/or chlorite. The breakdown of M2 kyanite to produce the M3 assemblage at < 0.5 GPa and 450–500 °C suggests a greenschist facies overprint during decompression. The P–T evolution of the FPM may represent subduction of an oceanic plateau with a granulate facies lower crust and subsequent exhumation in a Pacific-type orogen.

Key words: Early Palaeozoic; epidote–amphibolite; kyanite; polyphase metamorphism; P–T path; staurolite.

INTRODUCTION

Amphibolites are common in many Barrovian-type metamorphic terranes. Most contain higher-variance mineral assemblages that do not constrain well the metamorphic conditions. However, some high-pressure (HP) amphibolites with an aluminoous bulk-rock composition contain kyanite and staurolite (e.g. Gibson, 1979; Selverstone et al., 1984; Yokoyama & Goto, 1987; Gil Ibarguchi et al., 1991; Kuyumjian, 1998). Such unusual amphibolite assemblages are useful to constrain the P–T path (e.g. Arnold et al., 2000) as reaction textures are well displayed in aluminous minerals such as kyanite (e.g. Spear & Franz, 1986; Cotkin et al., 1988).

Early Palaeozoic kyanite- and staurolite-bearing epidote–amphibolite and metagabbro of the Fuku Pass metacumulate unit (FPM) occur in the peridotite body of Early Palaeozoic Oeyama belt (Kurokawa, 1975; Kuroda et al., 1976; Tsujimori, 1999). The Ky–St-bearing rocks have been considered to have recrystallized at medium-P (c. 0.5 GPa), possibly in an ocean-floor setting, as a cumulate member of an ophiolitic succession (Kurokawa, 1985). On the other hand, the Hbl + Czo + Ky + Pg + Rt assemblage suggests HP metamorphism (Tsujimori, 1999). Moreover, medium-P granulite-facies relics of Al-rich diopside and pseudomorphs of spinel and plagioclase were recently identified in some metacumulate rocks (Tsujimori & Ishiwatari, 2002). These rocks constrain a unique metamorphic evolution from the granulate-facies to HP metamorphism. The occurrence of an Early Palaeozoic HP metamorphic event in the FPM may signify the earliest subduction-metamorphism in the Pacific-type orogen of the Japanese Islands.

This paper presents new petrological data for the Ky–St-bearing epidote–amphibolite and metagabbros of the FPM, which are coupled with previous data to evaluate a polyphase metamorphic evolution of the FPM and discuss the tectonic significance of this evolution. Mineral abbreviations are after Kretz (1983) and the term ‘hornblende (Hbl)’ is used to describe Ca-amphibole with dominantly paragentic, tschermakitic and edenitic composition throughout this paper.
GEOLOGICAL SETTING

The Oeyama belt is an Early Palaeozoic ophiolitic nappe that occupies the highest structural position in the Phanerozoic Pacific-type orogen of south-west Japan (e.g. Ishiwatari & Tsujimori, 2003) (Fig. 1). Several serpentinitized peridotite bodies in the Chugoku Mountains consist mainly of residual peridotite with minor gabbro dykes and podiform chromitites (Arai, 1980; Kurokawa, 1985; Matsumoto et al., 1997). The peridotite has a lherzolitic composition in the eastern part, but is harzburgitic in the western part. The gabbro dykes have MORB-like affinity and yield Sm/Nd isochron ages of c. 560 Ma (Hayasaka et al., 1995), suggesting a Cambrian or earlier age for the ophiolite formation. The Oeyama peridotite may have been derived from suprasubduction zone mantle beneath an intraoceanic arc (Tsujimori & Itaya, 1999; Ishiwatari & Tsujimori, 2003).

Two different HP mafic rocks are associated with serpentinitized peridotite bodies in the Oeyama belt. The younger HP rocks consist of 330–280 Ma blueschist to eclogite facies metasediments and minor metabasalt, and represent fragments of the Late Palaeozoic Renge blueschist nappe underlying the Oeyama belt (Nishimura, 1998; Tsujimori, 1998; Tsujimori & Itaya, 1999). In contrast, the older HP rocks are characterized by 470–400 Ma deformed metagabbro and metaclinopyroxenite (Nishimura & Shibata, 1989; Tsujimori et al., 2000). The FPM belongs to the latter type.

The FPM occurs as a fault-bounded slice at the topographically higher portion of the Oeyama peridotite body that tectonically overlies the Permian accretionary complex of the Akiyoshi belt (Fig. 1). The FPM is a meta-ultramafic–mafic complex consisting of metamorphosed clinopyroxenite, wehrlite and cumulate gabbro (Uda, 1984; Kurokawa, 1985). On the other hand, the Oeyama peridotite body is composed of fertile harzburgite with minor dolerite and gabbro dykes (Uda, 1984; Kurokawa, 1985). Both have been highly serpentinitized, and subsequently significantly overprinted by contact metamorphism around Late Cretaceous granitic intrusions. Uda (1984) identified five contact metamorphic zones: I – Atg; II – Ol + Atg + Di; III – Ol + Atg + Tr; IV –Ol + Tlc ± Tr ± Cum; and V – Ol + En ± Tr ± Hbl. Primary igneous phases, Ol

Fig. 1. (a) A simplified map of the Chugoku Mountains, showing various pre-Triassic petrotectonic units and the locality of the Oeyama area. (b) Geological map of the Oeyama area, showing sample locality of the Ky–St-bearing rocks (after Tsujimori et al., 2000).
(F_{O_{99.1}}) + Opx (Al_2O_3 = 1.2–4.3 wt.%) + Cpx (Na_2O < 0.3 wt.%) + Cr-Spl (Cr# = 0.34–0.38), are well preserved in Zone I rocks (Kurokawa, 1985; Tsujimori, 1999). The FPM extends from Zone II to Zone V; few outcrops lack the thermal effect of the rocks (Kurokawa, 1985; Tsujimori, 1999). The FPM extends from Zone II to Zone V; few outcrops lack the thermal effect of the rocks (Kurokawa, 1985; Tsujimori, 1999). The FPM extends from Zone II to Zone V; few outcrops lack the thermal effect of the rocks (Kurokawa, 1985; Tsujimori, 1999).

Samples were collected at an outcrop (20 × 3 m) near the Fuko Pass where the effect of contact metamorphism is minimal (Kurokawa, 1985). Ky–St-bearing rocks occur together with irregular blocks of massive coarse-grained metagabbro, virtually undeformed, in the matrix of foliated (or gneissose) epidote–amphibolite. Tsujimori (1999) used the term ‘metagabbro’ to indicate any undeformed epidote–amphibolite in which original gabbroic textures are well-preserved, and distinguished it from the ‘foliated epidote–amphibolite’. The metagabbro is subdivided into leucocratic and melanocratic varieties by the content of clinzoisite.

**ANALYTICAL METHODS**

Concentrations of major (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K & P) and trace (Ni, Cu, Zn, Pb, Y & V) elements were analyzed by a Rigaku System 3270 X-ray fluorescence spectrometer with Rh tube at Kanazawa University. The operating conditions for both major and trace elements were 50 kV accelerating voltage and 20 mA beam current. Other trace elements (Sc, Cr, Co, La, Sm, Eu, Lu & Th) were determined by instrumental neutron activation analysis (INAA method). The INAA samples were activated at Kyoto University Reactor, and the gamma-ray spectroscopic analyses were done at the Radioisotope Laboratory of Kanazawa University.

Electron microprobe analysis was carried out with a JEOL JXA-8800R at Kanazawa University and JEOL JXA-8900R at Okayama University of Science. The quantitative analyses of rock-forming minerals were performed with 15 kV accelerating voltage, 12 nA beam current and 3–5 μm beam size. Natural and synthetic silicates and oxides were used for calibration. The ZAF method (oxide basis) was employed for matrix corrections.

**PETROGRAPHY**

Three rock types are distinguished according to their lithological and petrographical features: (1) foliated epidote–amphibolite (FEA); (2) leucocratic metagabbro (LMG); and (3) melanocratic metagabbro (MMG). Although epidote–amphibolite facies assemblages are dominant in all lithologies, each rock type preserves textual evidence for mineral growth during an earlier stage and of retrograde metamorphic events. Three metamorphic stages, M1, M2 and M3, are distinguished on the basis of microtextural relationships and mineral chemistry. Mineral parageneses for different metamorphic stages of these rock types are summarized in Fig. 2. M2 defines the peak HP metamorphism of all rock types, whereas M1 minerals are identified as relict minerals, particularly granulite facies relics and pseudomorphs in melanocratic metagabbro. M3 represents lower-P retrogression during decompression.

**FEA: foliated epidote–amphibolite**

This rock type consists mainly of hornblende, clinzoisite and kyanite with small amounts of paragonite, rutile, chlorite, margarite, staurolite, corundum, zoisite, and rare albite and muscovite. Ilmenite, sulphides and apatite occur as accessories. A foliation defined by preferred orientation of nematoblastic hornblende, kyanite and paragonite is developed. Mosaic aggregates of clinzoisite are intergrown with paragonite, kyanite, rutile, hornblende and staurolite; these minerals also occur as inclusions in clinzoisite (Fig. 3a,b). Rare zoisite blebs (< 0.1 mm), albite and muscovite are included in clinzoisite. Some staurolite (< 0.5 mm) occurs as aggregates (up to 5 mm) with minor corundum (< 0.1 mm) and rare magnetite; the aggregate is armoured by kyanite and clinzoisite at the margin (Fig. 3c). These matrix minerals and inclusions in clinzoisite are thought to be in textural equilibrium, and the mineral assemblage Hbl + Czo + Ky ± St + Pg ± Rt ± Ab ± Crn characterizes the peak M2. Some kyanite crystals contain plagioclase as tiny inclusions (< 0.03 mm) (Fig. 3d); this may represent a precursor of the peak of M2, possibly a remnant of the M1 stage. Rutile contains exsolution blebs

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**Fig. 2.** Mineral parageneses for three stages of metamorphic recrystallization for foliated epidote–amphibolite and leucocratic metagabbro (FEA & LMG), and melanocratic metagabbro (MMG).
of ilmenite; some are rimmed by thin ilmenite. Retrograde M3 white mica occurs as coronitic aggregates around M2 kyanite (Fig. 3d). Chlorite and tiny plagioclase (< 0.05 mm) occur rarely as M3 minerals with M3 margarite around M2 kyanite. M3 chlorite replaces M2 hornblende along the cleavage and staurolite in aggregates. The assemblage Mrg + Pg ± Ms + Chl ± Pl characterizes the M3 retrogression. In some samples, epidote overgrowths occur at interfaces between M2 clinzoisite and M3 hydros minerals. Epidote may be a product of M3 or later.

**LMG: leucocratic metagabbro**

This rock type consists mainly of clinzoisite and kyanite with minor hornblende, margarite, paragonite, staurolite, rutile, chlorite, zoisite and muscovite. Rare ilmenite and apatite occur as accessories. Regression in this rock is more obvious than in the FEA. Coarse-grained clinzoisite (up to 15 mm) contains all M2 minerals as inclusions. Some kyanite grains show pale-blue pleochroism. Staurolite includes trace M2 paragonite. Kyanite is pseudomorphed by M3 margarite and paragonite (Fig. 3e,f). M3 muscovite occurs with M3 margarite and paragonite (Fig. 4). M3 margarite is associated with M3 chlorite and rare M3 plagioclase.

**MMG: melanocratic metagabbro**

This rock type consists mainly of hornblende, clinopyroxene, corundum-magnetite symplectite, clinzoisite and chlorite with minor kyanite, margarite, rutile and paragonite. Clinopyroxene is a relict mineral of M1; it occurs as equigranular grains (3–6 mm) partly replaced by M2 hornblende (Fig. 3g). The Crn–Mag symplectite has equidimensional shape (1–3 mm), and consists mainly of intergrowths of corundum and graphic magnetite with minor Zn-rich spinel. Prismatic M3 epidote (<1 mm in length) was interpreted to be a pseudomorph after spinel of the M1 assemblage. Most kyanite inclusions in clinzoisite are replaced by M3 margarite. M3 chlorite replaces M2 hornblende, and occurs as chlorite-rich clots (3–5 mm), or aggregates surrounding the corundum-magnetite symplectite. Prismatic M3 epidote (< 1 mm in length) and tiny magnetite are randomly oriented and frequently associated with the chlorite clots.

**Bulk-rock chemistry**

One FEA, two LMG and two MMG samples were selected for geochemical study. Analyzed bulk compositions are listed in Table 1. Major element analyses show that all lithologies contain extremely low SiO₂ (38.7–41.1 wt.%) and high Al₂O₃ (18.9–25.5 wt.%), CaO (13.1–15.3 wt.%), and low Na₂O + K₂O (< 2.2 wt.%) with 48–66 mol.% normative anorthite and 16–28 mol.% olivine. The Mg/(Mg + Fe*) ratios range from 0.45 to 0.65 and increase in the order of FEA < LMG < MMG; Fe* = total Fe as Fe²⁺. The FEA and MMG are enriched in Sr (402–479 p.p.m.) whereas the MMG shows enrichments of Ni (97–99 p.p.m.), Sc (135–143 p.p.m.), and Zn (166–224 p.p.m.). The N-MORB normalized trace element abundances of the analysed samples of FEA, LMG and MMG for comparison, are shown in Fig. 4 for comparison. They are characterized by a high concentration of large ion lithophile (LIL) elements, and show similar patterns to those from meta-anorthosite of the Western Gneiss Region, Norway (Cotkin, 1997) and kyanite–staurolite-bearing rock from the Cabo Ortegal complex (NW Spain) (Gil Ibarguchi et al., 1991) where the REE contents of the analysed samples suggest a cumulative origin. The bulk rock composition and the calculated norm suggest an origin from a troctolitic- and anorhositic cumulate with abundant anorthosite-rich plagioclase.

**MINERAL CHEMISTRY**

Representative electron microprobe analysis of rock-forming minerals in the FPM are presented in Table 2.

**Ca-amphibole**

The structural formulae of amphibole are calculated based on O = 23 and the Fe²⁺/Fe³⁺ ratio was estimated on the basis of total cation = 13, excluding Ca, Na and K (Leake et al., 1997).
The MMG contains significantly lower M4-site ([M4]Na) than that in the order of FEA (0.59) < LMG (0.68) < MMG (0.73). Hornblende of wt.% TiO2, whereas actinolitic or tremolitic compositions are identified. X

M3 epidote of the FEA and MMG is characterized by distinctively high FFM and LMG, apparently due to different bulk rock composition. X

Table 1. Bulk-rock compositions of the FPM.

<table>
<thead>
<tr>
<th>Rock type</th>
<th>FEA</th>
<th>LMG</th>
<th>MMG</th>
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<tr>
<td>Major-element compositions (in wt %)</td>
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<tr>
<td>SiO2</td>
<td>41.08</td>
<td>39.51</td>
<td>39.37</td>
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<tr>
<td>TiO2</td>
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<td>0.65</td>
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<tr>
<td>Al2O3</td>
<td>20.63</td>
<td>25.46</td>
<td>23.97</td>
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<td>Fe2O3*</td>
<td>12.78</td>
<td>10.44</td>
<td>11.02</td>
</tr>
<tr>
<td>MnO</td>
<td>0.16</td>
<td>0.11</td>
<td>0.12</td>
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<tr>
<td>MgO</td>
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<td>4.38</td>
<td>5.03</td>
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<tr>
<td>CaO</td>
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<td>Na2O</td>
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<tr>
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<tr>
<td>Total</td>
<td>98.45</td>
<td>97.18</td>
<td>96.85</td>
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Trace-element compositions (in ppm)

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<th>Element</th>
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<tr>
<td>Sc</td>
<td>43.3</td>
<td>30.1</td>
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<td>V</td>
<td>536</td>
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<td>Cr</td>
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<td>Co</td>
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<td>Cu</td>
<td>123</td>
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<tr>
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<td>La</td>
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<td>1</td>
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<td>Eu</td>
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</tr>
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<td>Lu</td>
<td>0.3</td>
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</tr>
<tr>
<td>Th</td>
<td>1.0</td>
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<td>n.d.</td>
</tr>
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</table>

*Total Fe as Fe2O3.

‘hornblende’ of the studied rocks are tschermakitic to parargitic compositions containing up to 18 wt.% Al2O3, 3.4 wt.% Na2O and 1.2 wt.% TiO2, whereas actinolitic or tremolitic compositions are identified at the rim of relatively coarse-grained crystals in the MMG (Fig. 5). Individual hornblende grains in the FEA and LMG are unzoned, but their compositions are variable at the thin-section scale. Minimum XMg (= Mg/(Mg + Fe2+)) values become higher in the order of FEA (0.59) < LMG (0.68) < MMG (0.73). Hornblende of the MMG contains significantly lower M4-site ([M4]Na) than that in the FFM and LMG, apparently due to different bulk rock composition.

Epidote group minerals
The XFe2+ = [Fe3+/(Fe3+ + Al)] of M2 clinozoisite in all lithologies varies from 0.10 to 0.24. The XFe2+ of a zoisite bleb is 0.03–0.07. M3 epidote of the FEA and MMG is characterized by distinctly high XFe2+ (0.26–0.29).

Kyanite
Pale-blue coloured kyanite in the LMG contains up to 1.3 wt.% Fe2O3, whereas other colourless kyanite has < 0.8 wt.% Fe2O3.

White mica
Chemical compositions of white mica are shown in Na(Pg)–K(Ms)–Ca(Mrg) diagrams of Fig. 6. M2 paragonite of the FEA contains slightly higher Mrg component (Mrg = 11–18 mol%) than that of the LMG and MMG (Mrg = 4–6 mol%). M2 paragonite inclusions in staurolite of the LMG are characterized by extremely high Mrg component (Mrg = 36–47 mol%). M3 muscovite of the FEA is characterized by 6.2–6.3 Si p.f.u. (O = 22) and 17–24 mol% Pg. M3 paragonite of the FEA and LMG contains 2–14 mol% Mrg. M3 margarite shows a wide compositional range; Pg content varies from 7 to 34 mol%. M3 muscovite is characterized by 6.0–6.2 Si p.f.u. (O = 22) and < 10 mol% Pg.

Staurolite
The structural formulae of staurolite were calculated based on O = 46. Staurolite of the FEA contains a slightly higher XMg (0.19–0.36) but lower ZnO (0.7–1.6 wt.%) than that of the LMG (XMg = 0.16–0.29, ZnO = 0.9–2.0 wt.%). The analysed staurolite shows good negative correlation of (Fe+Mg+Zn) against (Si+Ti+Al) (Fig. 7). The correlation overlaps a substitution line of (Mg + Fe3+)1.5xAl18-xSi8O46 that represents magnesian staurolite (up to 0.61 XMg) in corundum-bearing amphibolite from an eclogitic metagabbro unit of the Sambagawa belt, SW Japan (Yokoyama & Goto, 1987).

Feldspar
M1 plagioclase inclusion in M2 kyanite of the FEA contains 26–34 mol% An. M2 albite in clinozoisite of the EFA contains < 5 mol% An (maximum 12 mol%). M3 plagioclase varies from 18 to 38 mol% An.

Clinopyroxene
M1 clinopyroxene of the MMG is Al-rich diopside that contains up to 8.5 wt.% Al2O3, up to 0.9 wt.% TiO2 and < 0.34 wt.% Na2O. The XMg varies from 0.78 to 0.94. The Al content decreases gradually toward the contact with clinozoisite.

Chlorite
M3 chlorite of the FEA and LMG is characterized by low Si (5.3–5.5 p.f.u. for O = 28) and high Al (5.1–5.8 p.f.u.) XMg = 0.71–0.77 in the FEA, and 0.64–0.77 in the LMG. In the MMG, M3 chlorite around the corundum–magnetite symplectite is aluminium (Al p.f.u. = 5.6–5.9, O = 28) with XMg ranging from 0.70 to 0.75, whereas that in chlorite-rich clots shows a wide compositional range (Al p.f.u. = 5.0–5.9, XMg = 0.65–0.67).

Others
Rutile contains 0.3–0.8 wt.% Fe2O3. Corundum in staurolite aggregates of the FEA contains 0.8–1.2 wt.% Fe2O3. Corundum in the corundum-magnetite symplectite of the MMG contains 0.6–1.7 wt.% Fe2O3 and < 0.1 wt.% Cr2O3. Magnetite intergrown with the corundum contains 0.2–1.4 wt.% Al2O3 and 0.2–0.4 wt.% TiO2. Zn-bearing spinel commonly shows exsolution texture; Ti-rich spinel lamellae (0.5–6.0 wt.% TiO2, 7–14 wt.% ZnO and XMg = 0.24–0.51) are developed parallel to the [100] plane of Ti-poor spinel (< 0.2 wt.% TiO2, 10–21 wt.% ZnO and XMg = 0.43–0.92). Ilmenite in the corundum-magnetite symplectite contains 2–7 wt.% MnO, whereas that in the FEA and LMG contains < 2 wt.% MnO.

METAMORPHIC CONDITIONS
Based on the observed petrographic features, a counter clockwise P–T evolution that passes from a medium-P granulite facies stage (M1) through a HP metamorphic
Table 2. Representative electron-microprobe analyses of rock-forming minerals in the FPM.

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*Total Fe as FeO₂.
**Total Fe as FeO.
peak (M2) to retrogression (M3) was identified from the kyanite–staurolite-bearing rocks in the FPM (Fig. 2). No Fe–Mg exchange geothermobarometer is applicable for the observed mineral assemblage. Hence, metamorphic $P$–$T$ conditions for each metamorphic stage are estimated based on available petrogenetic grids and phase equilibria. In this study,

![Fig. 5. Chemical compositions of Ca-amphibole on various plots (see text for details).](image)

![Fig. 6. K-Na-Ca plots of compositions of white micas from M2 and M3 stages. Compositional ranges of white micas by Frey et al. (1982), Guidotti, 1984 and Grapes & Palmer (1996) are also illustrated.](image)

![Fig. 7. Compositional variation of analyzed staurolite from FEA and LMG. The dashed line represents a substitution line (Mg + Fe)$_{3-1.5}$Al$_{1.5}$Si$_6$O$_{24}$ of magnesian staurolite from the Sambagawa belt (Yokoyama & Goto, 1987).](image)
calculations to obtain phase equilibria were carried out using version 1.1 of the software BAYES with an internally consistent thermodynamic dataset (Chatterjee et al., 1998).

M1: relict granulite-facies stage

The M1 assemblage is preserved either as relict Al-rich diopside or as pseudomorphs after spinel and plagioclase. Al-rich diopside is an index mineral in lower crustal basic and ultrabasic granulites with the Cpx + Opx + Pl + Spl assemblage (e.g. Rivalenti et al., 1981; Ishiwatari, 1985; Wilshire et al., 1991; McGuire, 1994). Although no relict plagioclase occurs in the MMG, the presence of relict plagioclase within M2 kyanite of the FEA suggests its occurrence prior to kyanite crystallization. Orthopyroxene may also have originally been present in the M1, perhaps having been consumed in subsequent metamorphic overprints. The Al-rich Cpx + Spl (pseudomorph) + Pl (pseudomorph) assemblage suggests the medium-P granulite facies (0.8–1.3 GPa at > 800 °C), which is bracketed by the spinel–gabbro field in the system CaO-MgO-Al2O3 (Gasparik, 1984; Schmädicker, 2000). In the spinel–gabbro field, the Al content (Ca-tschermakite component) of clinopyroxene is higher than that in the olivine–gabbro and garnet–gabbro fields, and is strongly T-dependent (e.g. Obata, 1976; Gasparik, 1984). The Al content (up to 8.5 wt.% Al2O3) of relict clinopyroxene suggests a temperature of around 800–900 °C.

M2: kyanite–staurolite-bearing HP stage

The kyanite–staurolite-bearing M2 assemblage, Hbl + Czo + Ky ± St + Pg + Rt ± Ab ± Crn, characterizes the peak HP metamorphism of all FPM rock types. Since the FEA and MMG contain cordum instead of quartz, the equilibrium conditions were described by a model system projected from cordum. In the model CASH (CaO-Al2O3-SiO2-H2O) system with excess Crn + H2O, the P–T limit of the Ky + Czo stability is defined by the following reactions (Fig. 8):

\[ \text{Dsp} = \text{Crn} + \text{H}_2\text{O} \]  
(1)

\[ 4\text{An} + \text{Crn} + \text{H}_2\text{O} = 2\text{Ky} + 2\text{Czo} \]  
(2)

and

\[ 4\text{Mrg} = 2\text{Ky} + 2\text{Czo} + 3\text{Crn} + 3\text{H}_2\text{O} \]  
(3)

The presence of paragonite, staurolite and albite provides more P–T constraints; the stability fields of these phases are bounded by the NASH (Na2O-ASH) and FASH (FeO-ASH) reactions listed below:

\[ \text{Ab} + \text{Crn} + \text{H}_2\text{O} = \text{Pg} \]  
(4)

\[ Jd_{50} (\text{Omp}) + \text{Ky} + \text{H}_2\text{O} = \text{Pg} \]  
(5)

\[ 2\text{Fe-Cld} + 4\text{Ky} + \text{Crn} = \text{Fe-St} + 3\text{H}_2\text{O} \]  
(6)

and

\[ Jd + \text{Qtz} = \text{Ab} \]  
(7)

The right hand sides of these reactions are stable in the M2. The stability fields shift towards the lower-P side and lower-T side with decreasing activities of clinozoisite and staurolite, respectively (Fig. 8). The P–T condition calculated using the software BAYES (Chatterjee et al., 1998) with activities of minerals listed in Table 3 is 600–850 °C and 0.95–1.90 GPa.

On the other hand, inclusion of relict plagioclase in M2 kyanite has probably remained from the reaction (2) during cooling with hydration; this reaction constrains a P–T limit for the M2 to 550–800 °C and 1.1–1.9 GPa according to the BAYES calculation (Fig. 8). The compositional gap between M2 paragonite and M2 muscovite yields a temperature of c. 600 °C (Blencoe et al., 1994). The Mrg-rich paragonite within staurolite may suggest a temperature of > 600 °C (Franz, 1977). The high Al (up to 18 wt.% Al2O3) and moderate Na (up to 3.4 wt.%) contents of hornblende further support the HP condition for these rocks (Ernst & Liu, 1998; Niida & Green, 1999).

M3: decomposition stage

The M3 is documented by minor retrograde coronas armoured around M2 kyanite, and is characterized by the Mrg + Pg ± Pl + Chl assemblage. This assemblage is a breakdown product of Ky + Czo according to the following reactions (Fig. 8):

\[ 2\text{Ky} + 2\text{Czo} = \text{Mrg} + 3\text{An} \]  
(8)

and

\[ 2\text{Ky} + 2\text{Czo} + \text{Ab} = \text{Pg} + 4\text{An} \]  
(9)

As both reactions (8) and (9) have gently positive P–T slopes, the M3 assemblage must involve a component of decompression; its P–T trajectory must cross these reaction curves. Staurolite was not replaced by the Cld + Ky assemblage during retrogression; hence reaction (6) also constrains the lower temperature limit to 450–500 °C at a maximum pressure of < 0.5 GPa for the M3 using the BAYES calculation (Fig. 8). The Pg–Mrg gap of the M3 mica suggests a temperature of c. 480 °C (Blencoe et al., 1994).

DISCUSSION AND CONCLUSION

The FPM rocks preserve minor relict middle-P granulite facies (M1) assemblage of Cpx + Pl + Spl ± possible Opx. They have significantly recrystallized under the HP epidote–amphibolite facies metamorphism (M2) close to the eclogite facies and formed the Hbl + Czo + Ky ± St + Pg + Rt ± Ab ± Crn assemblage (M2). Subsequent decompression produced the Mrg + Pg ± Ms ± Pl + Chl.
assemblage (M3). The \( P-T \) path for these three metamorphic events is shown in Fig. 8.

The estimated \( P-T \) conditions of the \( M_2 \) imply a geothermal gradient around 15° C km\(^{-1}\), indicating a subduction zone metamorphism. The decompression path from \( M_2 \) to \( M_3 \) is roughly isothermal and similar to some collision-type HP-UHP rocks (e.g. Maruyama et al., 1996). The \( M_3 \) may represent a greenschist facies overprint during exhumation; hornblende K/Ar ages of 443–403 Ma (Tsujimori et al., 2000) may signify the timing of exhumation during the trajectory \( M_2-M_3 \). The timing of the \( M_1 \) has not been dated, but it must be older than the K/Ar ages.

The presence of medium-\( P \) granulite-facies (M1) and the cooling trajectory to the HP metamorphism (M2) is an unusual feature, particularly in a Pacific-type orogen. Lower crustal granulites that were recrystallized and hydrated under HP or UHP metamorphism occur in many continental collision zones (e.g. Gil Ibarguchi et al., 1991; Tenthorey et al., 1996; Münttener et al., 2000), because granulite facies rocks constitute major Precambrian continental crust. However, subducted ‘allochthonous’ granulites are less common in Pacific-type orogens. The relict granulite facies assemblage (Cpx + Opx + Pl + Sp + Grt) in the eclogite unit of Cretaceous Sambagawa belt, Japan is a rare example. There, the relict granulites of basic to ultrabasic cumulate origin were recrystallized under the Cretaceous Sambagawa HP metamorphism to form unusual assemblages of Ky + Zo + Grt and rare Ky + St + Crn (Yokoyama, 1980; Yokoyama & Goto, 1987). The metamorphic evolution of the FPM described above is similar to that of the Sambagawa ‘granulites’.

Medium-\( P \) granulite facies conditions are attained at Moho depth beneath large oceanic plateaux (e.g. Saunders et al., 1996). In fact, rare oceanic two-pyroxene granulites have been accreted as a minor component of ophiolite to the circum-Pacific orogen (e.g. Yakuno (SW Japan); Ishiwatari (1985); Tonshina (Alaska); Debari & Coleman (1989); Bikin (Far East Russia); Vysotskii, 1994); they have also been described as lower crustal xenoliths in the present-day oceanic plateau (Gregoire et al., 1994). The bulk-rock chemistry implies
that the protoliths of the Ky-St-bearing rocks in the FPM are correlative with troctolitic cumulus rocks, and coincident with cumulates of the ‘non-island arc origin’ ophiolites (Tsujimori & Ishiwatari, 2002). The relic granulite facies in both FPM and Sambagawa may provide constraints on the subduction of thick oceanic crust. Saunders et al. (1996) argued that the introduction of H_2O-rich fluid at the lower crustal part of an oceanic plateau allows transformation of gabbroic granulite to eclogite. They pointed out that the collision of thick oceanic crust causes a backward migration of the subduction zone, and consequently H_2O-rich fluid released from the subducting slab promotes transformation of suprasubduction zone ophiolitic rocks to eclogite. The unusual P–T evolution of the FPM may represent subduction of a thick-crustal oceanic plateau and subsequent exhumation.

Early Palaeozoic HP metamorphic rocks of the Kurosegawa belt (Maruyama & Ueda, 1974) and the Oeyama belt (Tsujimori et al., 2000) provide a petrotectonic constraint for the earliest subduction event in the Japanese orogen. The HP rocks described in this paper occur in an early Palaeozoic subduction zone with geothermal gradient in the order of 15 °C km⁻¹. Such a relatively high geothermal gradient in the subduction zone has produced epidote-amphibolite facies metamorphic rocks. On the other hand, typical blueschists-eclogites with a low-geothermal gradient around 10 °C km⁻¹ have been exhumed only after Devonian-Carboniferous time during a continuous subduction of colder oceanic lithosphere (Ueda et al., 1980; Tsujimori & Itaya, 1999; Tsujimori, 2002). Such a difference and many other examples documented elsewhere led to a suggestion of secular cooling of Earth and subduction-zone geotherm by Maruyama et al. (1996).

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