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Tectono-metamorphic evolution of high-P/T and low-P/T metamorphic rocks in the Tia Complex, southern New England Fold Belt, eastern Australia: Insights from K–Ar chronology

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ABSTRACT

The Tia Complex in the southern New England Fold Belt is a poly-metamorphosed Late Paleozoic accretionary complex. It consists mainly of high-P/low-T type pumpellyite–actinolite facies (rare blueschist facies) schists, phyllite and serpentinite ($T = 300$ °C and $P = 5$ kbar), and low-P/high-T type amphibolite facies schist and gneiss ($T = 600$ °C and $P < 5$ kbar) associated with granodioritic plutons (Tia granodiorite). White mica and biotite K–Ar ages distinguish Carboniferous subduction zone metamorphism and Permian granitic intrusions, respectively. The systematic K–Ar age mapping along a N–S traverse of the Tia Complex exhibits a gradual change. The white mica ages become younger from the lowest-grade zone (339 Ma) to the highest-grade zone (259 Ma). In contrast, Si content of muscovite changes drastically only in the highest-grade zone. The regional changes of white mica K–Ar ages and chemical compositions of micas indicate argon depletion from precursor high-P/low-T type phengitic white mica during the thermal overprinting and recrystallization by granitoids intrusions. Our new K–Ar ages and available geological data postulate a model of the eastward rollback of a subduction zone in Early Permian. The eastward shift of a subduction zone system and subsequent magmatic activities of high-Mg andesite and adakite might explain formation of S-type granitoids (Hillgrove suite) and coeval low-P/high-T type metamorphism in the Tia Complex.

1. Introduction

The New England Fold Belt (NEFB) of Western Australia is a Pacific-type accretionary orogenic system with oceanward growth since Early Paleozoic time, as supported by considerable field-geologic and biostratigraphic data. The growth of NEFB is characterized by a series of orogen-parallel accretionary complexes containing syn or post-orogenic calc-alkaline plutons; each accretionary complex is characterized by an oceanic plate stratigraphy. In Pacific-type orogen, granitoid magmatism accompanied with low-P/high-T (low-P/T) type metamorphism enhances growth and modification of continental arc crust (e.g., Miyashiro, 1961; Maruyama, 1997). Therefore geochronological studies of the low-P/T type metamorphic domains are important to understand tectono-metamorphic evolution of the middle crust.

The NEFB is mainly composed of the subduction–accretion complexes and the forearc basin sediments (Scheibner and Basden, 1996; Cawood et al., 2011). Systematic K–Ar dating of metamorphic rocks from the southern part of the NEFB has indicated the existence of three metamorphic episodes, ca. 260 Ma, ca. 330 Ma and ca. 470 Ma (Fukui, 1997; Fukui and Itaya, 1990; Fukui et al., 1990, 1993, 1995). The older metamorphic events of ca. 330 Ma and ca. 470 Ma were recorded in the high-P/low-T (high-P/T) type blueschists-bearing rocks. The younger metamorphic event of ca. 260 Ma was recorded in the low-P/high-T type gneisses occurring in the Tia Complex. These two contrasting rock types (high-P/T and low-P/T) in the Tia Complex have been discussed by Gunthorpe (1970) and Dirks et al. (1992). In particular, Gunthorpe (1970) suggested that the low-P/T assemblages resulted from emplacement of granodioritic intrusions, and subsequently the complex had buried to form the blueschist. However, his model is contradictory to the fact that low-P/T gneisses are significantly younger than the high-P/T blueschists. On the other hand, Dirks et al. (1992) interpreted a thermal overprint for a metamorphosed accretionary complex in a compressional back-arc setting. In either case, the tectonic relationship between low-P/T type metamorphic

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[1] Obituary: One of the authors (Teruo Watanabe) fell down accidentally from a cliff during a field survey in the Toyohama Tunnel landslide site of western Hokkaido, Japan and ended his life on 9 May, 2002.

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rocks and associated granitoids is still debated. The systematic K–Ar age mapping along a metamorphic gradient that characterizes the relationship between temperature and deformation/recrystallization in a single metamorphic sequence provided critical information on tectono-metamorphic processes (e.g., Itaya and Takasugi, 1988; Nishimura et al., 2000). As we presented in this paper, we have applied K–Ar age mapping method for the low-P/T domain within a high-P/T metamorphic sequence in the Tia Complex of the southern NEFB. We also describe regional changes of mineralogical properties of biotite, white micas and carbonaceous materials in metasediments. Combined with the available geologic data in the Tia Complex and new K–Ar dating, we discuss the tectono-metamorphic evolution of the Tia Complex.

2. Outline of geology

The New England Fold Belt is mainly composed of the subduction–accretion complexes and the fore arc basin sediments (e.g., Scheibner and Basden, 1996). In the southern NEFB, the forearc basin sediments belong to the Tamworth belt and the subduction–accretion associations, to the Tablelands complex (Korsch, 1977; Cawood and Leitch, 1985; Cross et al., 1987). The boundary between the Tamworth belt and the Tablelands complex is the Peel Fault along which the Woodsreef serpentinite mélange exposes (Blake and Murchey, 1988) and contains middle Ordovician high-P/T metamorphic rocks (Fukui et al., 1995). The high-P/T metamorphic rocks such as blueschists are also found in the Tia Complex and adjacent area in the Tablelands complex (Gunthorpe, 1970; Watanabe and Hirama, 1988a; Fukui et al., 1993). The Tia Complex is composed of a wide range of metamorphic rocks (such as gneiss, schist and phillite of pumpellite–actinolite to amphibolite facies), which have been multiply deformed and poly-metamorphosed due to the Tia granodiorite intrusion (Gunthorpe, 1970; Watanabe and Naka, 1988). Rock types in the Tia Complex are siliceous schist, metagreywacke, metasiltstone, slate, pebbly metapelite, metabasite, chert and serpentintie. Gunthorpe (1970) divided the Complex into the more basic and chert-rich Oxley and Wybeena metamorphics and the pelitic Brackendale metamorphics (Fig. 1). The former two were considered by Korsch (1977) to belong to the Sandon Association, whereas the Brackendale metamorphics were interpreted as part of the Coffs Harbour Association. The Tia Granodiorite forms part of the Hillgrove Suite, which consists of S-type plutons with low initial $^{87}Sr/^{86}Sr$ ratios, that are interpreted to have been derived from volcanoclastic metasediments (Flood and Shaw, 1977; Shaw and Flood, 1981). Zircons from the granodiorite yield U–Pb age of 296 Ma (Cawood et al., 2011). The Tia Complex is bounded by a fault on the west and southwest margins with discontinuous serpentinite. The eastern margin is assigned to be a fault on the basis of the distribution of mafic rocks.

The Oxley metamorphics were characterized by glaucophane bearing schists (Gunthorpe, 1970; Watanabe and Hirama, 1988a). The sodic amphibole/winchite is also recognized in the area beyond the eastern boundary (Watanabe and Hirama, 1988a). Thus, the rocks of blueschist facies in a wide sense are considered to occur in a wide area of the Tablelands complex. Mafic rocks occur commonly in the southern part of the Oxley metamorphics. Pillow structure is preserved in a few places (Watanabe and Naka, 1988), suggesting the host lithologies of most mafic rocks are of hyaloclastic origin. The siliceous green schists associated with the mafic rocks are also of hyaloclastic origin; pelitic rocks (schist/phylite) are rarely interbeded in the mafic rocks. Psammitic and pelitic rocks are predominant in the Brackendale metamorphics. The main trend of these rocks is NW–SE with small-scaled NE–SW trending fold (Watanabe and Hirama, 1988b; Watanabe and Naka, 1988). Steeply dipping foli-

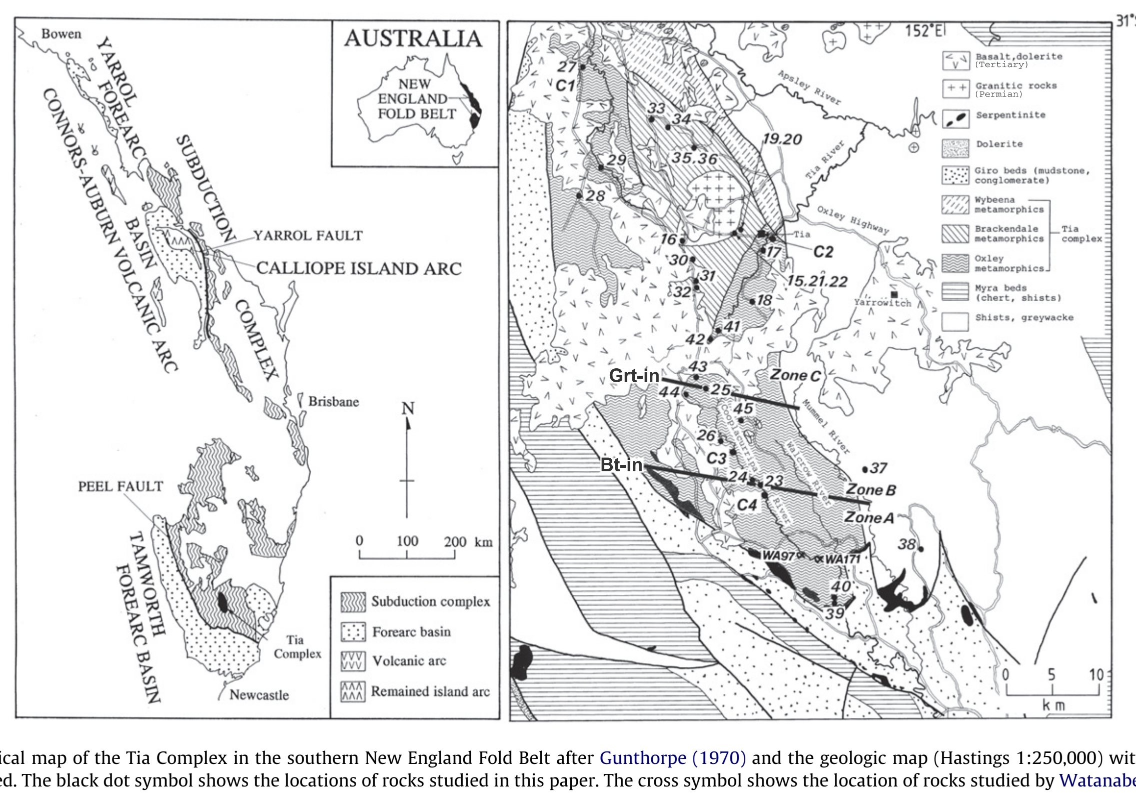

Fig. 1. Geological map of the Tia Complex in the southern New England Fold Belt after Gunthorpe (1970) and the geologic map (Hastings 1:250,000) with location of the samples studied. The black dot symbol shows the locations of rocks studied in this paper. The cross symbol shows the location of rocks studied by Watanabe et al. (1988a,b). Numbers C1 to C4 indicates the rocks of which only carbonaceous material analyses were carried out. For other samples with numerals, both K–Ar and carbonaceous analyses were carried out.

ation is characteristically recognized around the Tia granodiorite and multiple deformations are common.

Dirks et al. (1992) described a thermal overprint of a subduction–accretion complex in the Tia Complex and argued the P–T–t deformation path based on eight overprinting deformation events. They include the fabric-forming folding events F1–F5, a non-fabric-forming folding event F6, a ductile shearing event D7, and a brittle faulting event D8. Rocks in the Tia Complex moved from mid-crustal levels at blueschist facies conditions (Dl, D2; 6 kbar, 200 °C) to upper-crustal levels at high-T amphibolite facies conditions (D4, D5; 2.5 kbar, 600 °C), in a relatively short time (~15 Ma). The rocks remained under these conditions for approximately 40 Ma (D6) before they cooled and were brought to the surface (D7, D8). During D6, Permian rift basins were opening as thermal gradients in the Tia Complex reached a maximum (Dirks et al., 1992).

On the basis of index minerals in pelitic/psammitic rocks, Gunthorpe (1970) divided the area into three zones, A (biotite free zone), B (biotite bearing zone) and C (garnet bearing zone) as shown in Fig. 1. Watanabe and Hirama (1988a) recognized much wider occurrence of biotite in pelitic schists and the difference of first appearance of garnet in the pelitic and siliceous schists.

3. K–Ar and XRD analyses

We collected 35 rock samples from the Tia Complex to carry out K–Ar analyses of biotites and white micas and XRD analyses of carbonaceous material. The samples were crushed with a jaw crusher and then sieved. Biotites and white micas were separated from the several size fractions of the samples. We chose the best fraction for the abundance and purity and then carried out K and Ar analyses. Analysis of potassium and argon of biotite and 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 μg/g Cs buffer with an analytical error within 2% at a 2σ 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 ^{38}Ar spike. Multiple runs of the standard (JG-1 biotite, 91 Ma) indicate that the error of argon analysis is about 1% at a 2σ confidence level (Itaya et al., 1991). The decay constants of ^{40}K to ^{40}Ar, ^{40}Ca, and ^{40}K content in potassium used in the age calculations are 0.581×10^{-10} year^{-1}, 4.962×10^{-10} year^{-1} and 0.0001167, respectively (Steiger and Jager, 1977).

Carbonaceous material was separated from silicate minerals by the method similar to that used by Itaya (1981). A 50–150 g of pelitic rock was crushed in a stamp mill. Material passing through a 100 mesh screen was treated successively by (1) hydrochloric acid (1–2 N, 1–2 h), (2) hydrofluoric acid (55%, 8–10 h), and (3) hydrochloric acid (4–5 N, 4–6 h) on a hot plate. Step (2) and (3) were repeated at least twice. In these treatments, drying of the residue was avoided. Between acid treatments, the residues were washed with pure water and, at the end of the treatment, with ethanol. Insoluble minerals such as zircon, tourmaline and pyrite were removed by using difference of the sedimentation rate in ethanol. Treatment with nitric acid to remove pyrite was not attempted because carbonaceous material is easily oxidized (Saxby, 1970) and its crystal structure may be affected. This method yields 70–94% concentration of carbonaceous material. The carbonaceous material in ethanol was dried on a glass slide and analyzed with an X-ray diffractometer using Ni-filtered Cu Kα radiation. Diffractograms were calibrated with silicon standard.

K–Ar age data of biotite and white mica separates from the metamorphic rocks and the granodiorites and the apparent d_{002} values of carbonaceous materials from the metasediments in the Tia Complex are listed in Table 1. The potassium content indicates that the white mica fractions from some samples were probably impure. Impurities in the white mica fractions were generally quartz, albite and carbonaceous material. These impurities do not significantly affect the K–Ar ages (e.g., Itaya and Takasugi, 1988). Fig. 2 shows the K–Ar ages plotted along the N–S traverse from the center of the Tia granodiorite in the southern part of the complex. In the northern part of the complex, the ages are plotted at the distance from the center of the Tia granodiorite. The biotite ages of the Tia granodiorites gave 265 and 278 Ma that are within the biotite age variation (258–283 Ma) from the metamorphic rocks. The white mica ages show a systematic younging toward the granodiorite from the lowest-grade zone (No. 39; 339 Ma) far from the granodiorite to the highest grade zone close to the granodiorite (No. 19; 259 Ma).

Most low-grade zone metasediments of the high-pressure belts contain two types of white micas, the dusty and clear micas as observed in the southern part of the Sanbagwa belt in central Shikoku (Itaya and Fukui, 1994) and in the Piemonte calcschists of western Alps (Takeshita et al., 2004). The dusty micas are neoblastic and metamorphic origin; they stretch along a mica layer associated with carbonaceous material. The clear micas are detrital ones and occur as coarser grain. Some detrital micas are extremely elongated along the stretching lineation and are poorer in Si content than dusty metamorphic micas. Both psammitic and pelitic schists contain detrital micas derived from terrigenous clastics; these micas yield ages older than the expected ages. The lowest grade schist (No. 39) is the siliceous green schist of which the host lithology is of hyaloclastic origin and has no detrial micas. The age indicates the timing of exhumation of the high-P/T type of metamorphic rocks (cf. Itaya et al., 2011).

The samples 19, 20, 31 and 42 gave the biotite ages older than the white mica ages though the closure temperatures are lower in biotite than white mica. These older biotite ages may be due to the excess argon wave observed in a Barrovian type pelitic schist in the eastern Tibetan plateau, of which the host lithologies have experienced polymetamorphic events (Itaya et al., 2009).

4. Temperature and pressure conditions

Fig. 2 also shows the apparent d_{002} values plotted along the N–S traverse from the center of the Tia granodiorite in the southern part of the complex, indicating the temperature increases toward the Tia granodiorite. The d_{002} values provide an estimate of their metamorphic temperatures (Itaya, 1981). The d_{002} values smaller than 3.4 Å suggest that their metamorphic temperature is higher than 450 °C. The values larger than 3.5 Å suggest the pumpellyite–actinolite zone in the high-P/T type of metamorphic sequence (Itaya, 1981). Sodic amphibole, sodic pyroxene and pumpellyite are recognized in the mafic rocks in the southern part of the Oxley metamorphics (Gunthorpe, 1970; Watanabe and Hirama, 1988a; Watanabe, 1988), suggesting a low pressure part of the glaucophane schist facies. Judging from Al_2O_3 content (3%) in sodic amphibole and jadeite mole (17%) in sodic pyroxene coexisting with actinolite–albite–quartz, Watanabe (1988) estimated a condition of 4.5–5.0 kbar and 300 °C. While, Hand (1988) reported pseudomorph of lawsonite and estimate higher-P condition as the earliest metamorphic phase. These mineralogical signatures suggest the high-P/T of rocks formed in the subduction zone.

The mineral assemblage and limited migmatization of the metapelites at the granodiorite boundary suggest upper amphibolite grade with 600–650 °C. Some pelitic schists contain cordierite,

Table 1
K–Ar age data of white mica and biotite separates from the investigated metamorphic rocks and granodiorite of the Tia Complex. Mineralogical properties are also listed (See text).

Locality	Rock type	Mineral	Mesh size	K (wt%)		Rad. ^{40}Ar		Non rad. Ar	Age (Ma)		C.M.	White micas						Biotite	
					error	(10^{-8}ccSTP/g)	error	(%)		error	d_{002} (Å)	Si	stdev	X_{Mg}	stdev	Pg	stdev	Ti	stdev
27	pelitic	Ms	280-330	3.32	0.07	4401	61	0.4	312.8	7.0	3.53	3.43	0.10	0.53	0.09	1.5	2.0		
C1	pelitic										3.47								
28	pelitic	Ms	280-230	1.03	0.02	1286	18	0.5	296.2	6.6									
29	pelitic	Ms	280-330	4.39	0.09	5465	75	0.1	295.3	6.6		3.47	0.04	0.56	0.02	0.8	2.1		
33	psammitic	Ms	253-280	2.60	0.05	2769	39	0.2	255.4	5.8	3.37							0.120	0.016
34	pelitic	Ms	235-330	3.59	0.07	4291	59	0.4	284.4	6.4	3.37	3.15	0.06	0.46	0.03	4.0	0.4		
34	pelitic	Bt	235-330	6.54	0.13	7220	110	9.7	264.2	6.2								0.123	0.018
36	psammitic	Ms	150-200	6.47	0.13	7540	103	0.3	277.8	6.2	3.37								
34	psammitic	Bt	150-200	6.68	0.13	7760	115	8.2	277.0	6.4									
35	psammitic	Ms	253-280	6.16	0.12	7156	98	0.2	277.0	6.2	3.37	3.11	0.03	0.48	0.02	5.5	2.4		
35	psammitic	Bt	253-280	7.14	0.14	8248	122	7.7	275.5	6.4								0.155	0.019
16	granodiorite	Bt	80-100	7.65	0.15	8912	132	7.3	277.7	6.4									
15	granodiorite	Bt	80-100	7.04	0.14	7802	118	9.2	265.1	6.2									
22	pelitic	Bt	150-200	7.59	0.15	8363	84	1.9	263.6	5.5								0.151	0.012
C2	pelitic										3.37								
21	pelitic	Ms	150-200	6.54	0.13	7241	71	0.5	264.9	5.5									
21	pelitic	Bt	150-200	7.70	0.15	8505	88	0.7	264.3	5.5									
19	pelitic	Ms	150-200	3.74	0.08	4037	39	0.6	258.6	5.4		3.13	0.02	0.46	0.02	5.0	1.8		
19	pelitic	Bt	150-200	7.50	0.15	8499	84	1.9	270.8	5.6								0.155	0.014
20	psammitic	Ms	150-200	4.31	0.09	4754	47	0.6	264.1	5.5									
20	psammitic	Bt	150-200	7.48	0.15	8404	81	1.3	268.5	5.5									
30	psammitic	Ms										3.19	0.04	0.50	0.04	3.6	0.6		
30	psammitic	Bt	150-200	7.55	0.15	8797	127	5.9	277.7	6.4	3.36							0.148	0.016
17	pelitic	Ms	150-200	5.19	0.10	5817	57	1.1	267.7	5.5	3.37	3.18	0.05	0.48	0.06	4.8	2.4		
17	pelitic	Bt	150-200	6.81	0.14	7345	72	1.6	258.4	5.4								0.127	0.010
31	psammitic	Ms	235-280	3.54	0.07	3917	53	0.3	264.8	6.0	3.37	3.25	0.11	0.52	0.09	7.5	8.8		
31	psammitic	Bt	235-280	7.32	0.15	8369	125	8.1	272.9	6.3								0.140	0.010
32	psammitic	Ms	235-280	3.48	0.07	4109	57	0.2	281.2	6.3									
32	psammitic	Bt	235-280	6.81	0.14	7853	124	8.4	275.1	6.5									
18	pelitic	Ms	150-200	5.90	0.12	6477	63	0.8	263.0	5.4	3.37	3.13	0.05	0.44	0.04	5.3	4.6		
18	pelitic	Bt	150-200	6.88	0.14	7598	74	1.0	264.2	5.5								0.125	0.009
41	pelitic	Ms	150-200	7.90	0.16	9210	132	0.3	277.9	6.3	3.37	3.15	0.05	0.49	0.06	4.9	0.7		
41	pelitic	Bt	150-200	7.00	0.14	8071	128	9.4	275.0	6.5									
42	pelitic	Ms	150-200	7.90	0.16	9287	129	0.1	280.0	6.3	3.38								
42	pelitic	Bt	150-200	6.83	0.14	8113	119	7.4	282.7	6.5									
43	pelitic	Ms																0.124	0.012
43	pelitic	Bt	150-200	5.93	0.12	6490	101	11.8	262.0	6.2	3.38	3.39	0.09	0.62	0.05	2.2	0.4		
25	pelitic	Ms	200-235	7.42	0.37	8541	82	0.5	274.7	13.0		3.53	0.05	0.62	0.04	2.8	3.7		
25	pelitic	Bt																0.088	0.018
44	pelitic	Ms	150-200	5.70	0.11	7116	108	9.0	296.0	6.9		3.45	0.05	0.88	0.03	2.4	0.5		
45	pelitic	Ms	280-330	5.82	0.12	7015	106	9.3	286.6	6.7	3.38	3.42	0.09	0.59	0.06	2.4	1.2		
45	pelitic	Bt																0.100	0.017
26	pelitic	Ms	200-235	7.94	0.16	9958	93	0.7	297.4	6.1		3.50	0.02	0.47	0.02	0.6	0.3		
C3	pelitic										3.47								
24	pelitic	Ms	200-235	6.43	0.13	8389	87	0.5	308.2	6.4		3.49	0.02	0.70	0.01	0.8	0.2		
37	pelitic	Ms	330-400	2.64	0.05	3468	51	7.8	310.2	7.1									
23	pelitic	Ms	200-235	7.15	0.14	9251	89	0.5	305.9	6.2	3.51	3.43	0.06	0.49	0.02	2.0	2.1		
C4	pelitic										3.50								
38	mafic	Ms	235-330	6.56	0.13	8966	132	7.0	321.8	7.3		3.47	0.10	0.54	0.07	4.1	8.8		
40	mafic	Ms	330-400	4.64	0.09	6286	87	0.2	319.1	7.1									
39	mafic	Ms	235-330	6.53	0.13	9455	129	0.1	339.2	7.5		3.47	0.04	0.56	0.03	2.3	4.0		

Ms—White micas, Bt—biotite, CM—carbonaceous materials, Si—Si a.p.f.u. (O = 11), stdev—standard deviation, X_{Mg}—Mg/(Mg + Fe), Pg—100 × Na/(Na + K), Ti—Ti a.p.f.u. (O = 11).

suggesting the pressure is lower than 5 kbar. This also indicates that the rocks close to the Tia granodiorite had undergone low-P/T type metamorphism.

5. Chemistry of biotites and white micas

As mentioned above, genetically different types of metamorphic rocks occur in the Tia Complex; the high-P/T and the low-P/T types. To examine any relationships between the chemistries of minerals in both type of rocks, chemistry of white mica and biotite in the samples dated with K–Ar method mentioned above was analysed by an EPMA. Electron microprobe analyses of minerals were carried out at Department of Geology and Mineralogy, Hokkaido University using a JCMA Superprobe 733 MKII. The quantitative analyses on mineral chemistry were performed with 15 kV accelerating voltage, 20 nA beam current and 2–3 μm beam size. Natural and synthetic silicates and oxides were used for calibration. The ZAF method (oxide basis) was employed for matrix corrections.

The Si values, and Mg/(Mg + Fe) and Na/(Na + K) ratios of white micas, and Ti values of biotites are plotted along the N–S traverse from the center of the Tia granodiorite in the southern part of the complex and, in the northern part of the complex, at the distance from the center of the Tia granodiorite (Fig. 2). These figures indicate that the white micas in each specimen are heterogeneous, in particular, in the medium grade zone. White micas from the highest-grade zone (No. 19) are close to pure muscovite. On the other hand, white micas in the lowest-grade zone (No. 39) are pure phengite. It should be noted that the Si contents change drastically

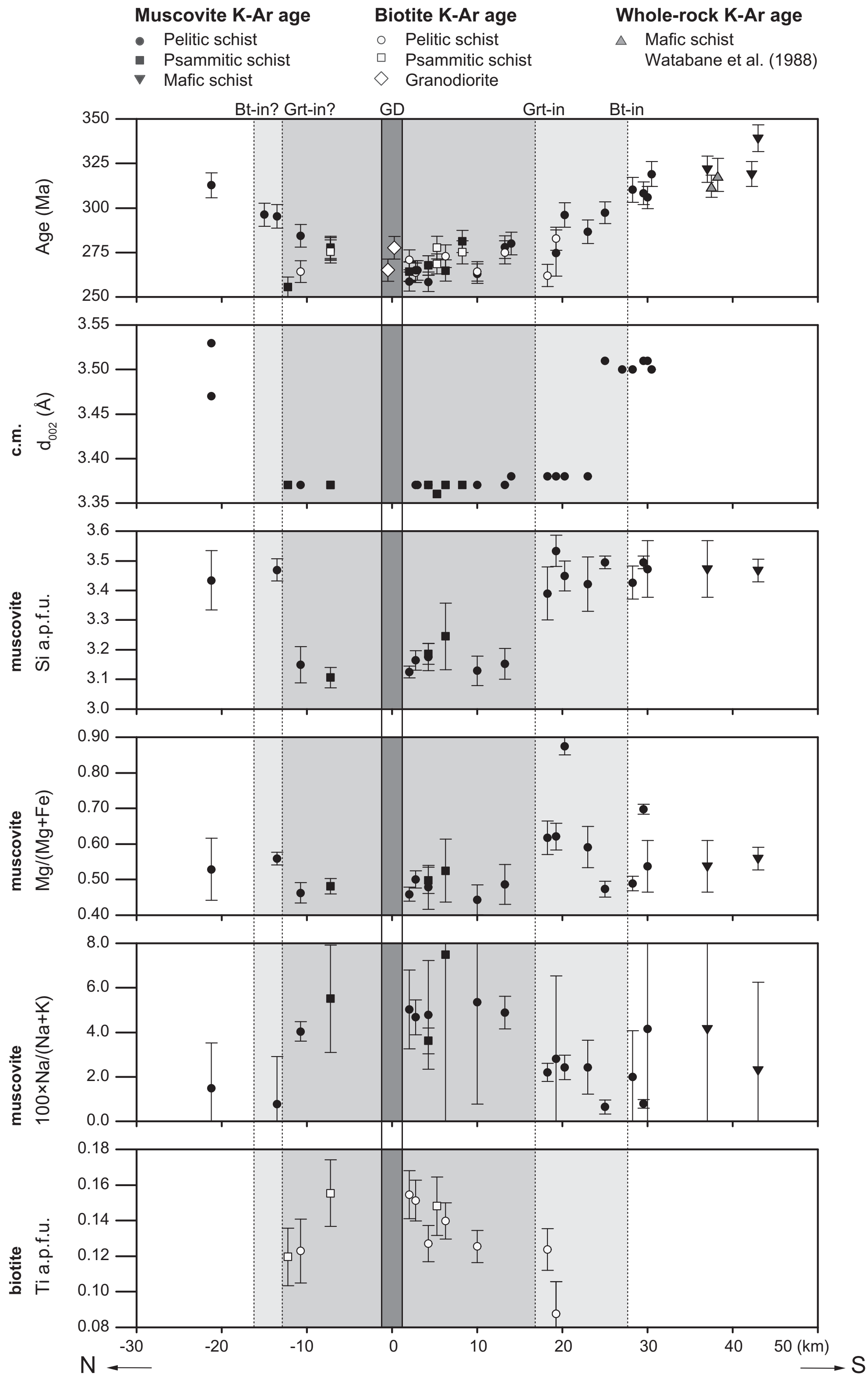

Fig. 2. K–Ar ages of white micas and biotite, and the apparent d_{002} values of carbonaceous materials are plotted along the N–S traverse from the center of the Tia granodiorite in the southern part of the Tia Complex. In the northern part of the complex, the values are plotted at the distance from the center of the Tia granodiorite. Compositional data of white micas and biotite are also plotted in the same manner.

at the garnet-in boundary. The Mg/(Mg + Fe) and Na/(Na + K) ratios are heterogeneous and have no significant difference between the high-P/T and low-P/T type rocks. The Ti contents in biotite increase with an increase in the metamorphic temperatures, strongly suggesting the biotites formed in equilibrium with other phases during the low-P/T type metamorphism.

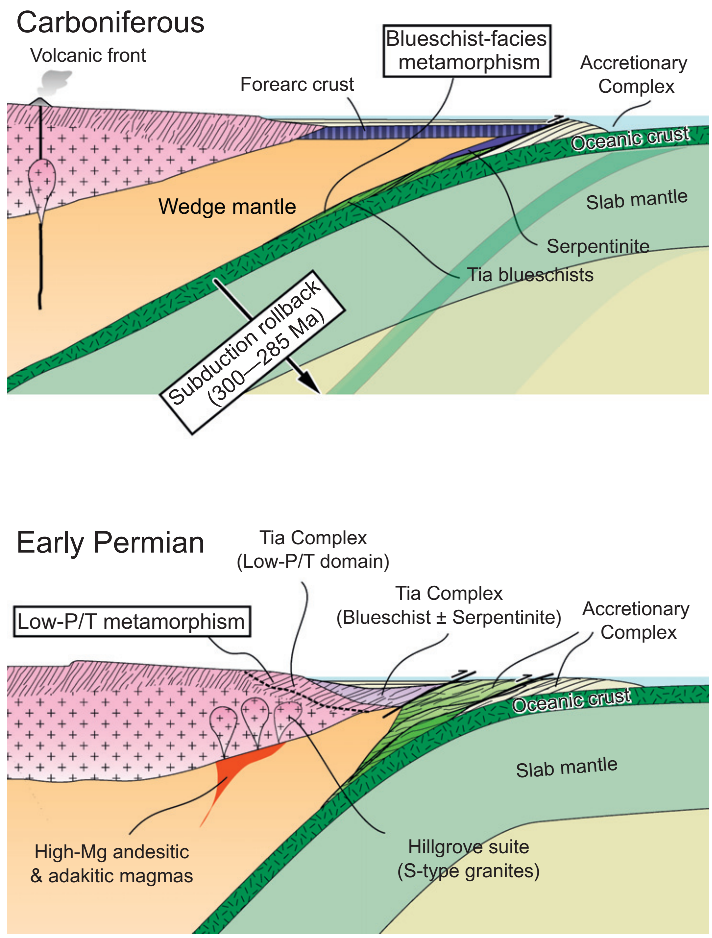

Fig. 3. A possible tectono-metamorphic evolution from Carboniferous to Early Permian for the Tia Complex. The eastward rollback of a subduction zone might play an important role for activities of high-Mg andesite and adakite to form S-type granitoids (see text). In the figure, the uppermost continental crust is composed of older accretionary complexes; this crustal structure model of continental margin is based on a recent seismic data of SW Japan (Ito et al., 2009).

6. Discussion

6.1. Argon depletion from white micas

The closure temperatures of K–Ar muscovite and biotite systems have been long believed to be 350 °C and 300 °C in metamorphic rocks with slow cooling rate, respectively (Purdy and Jäger, 1976). In the last decade, a number of geochronologists have cast doubt on the closure temperatures proposed by Purdy and Jäger (1976) (cf., Villa, 1998; Takeshita et al., 2004; Gouzu et al., 2006a,b; Itaya et al., 2009). Villa (1998) examined Jäger's calibration of the closure (or blocking) temperatures and proposed new closure temperatures of 500 °C for muscovite and 450 °C for biotite. Takeshita et al. (2004) studied the resetting temperature of detrital white micas based on the systematic K–Ar analyses of white micas in calcschists along the Chisone valley in the western Alps. Their results strongly support the new closure temperature proposed by Villa (1998). Itaya et al. (2009) carried out $^{40}Ar/^{39}Ar$ analyses of biotite and muscovite in a Barrovian type pelitic schist in the eastern Tibetan plateau, of which the host lithologies have experienced polymetamorphism as suggested by multichronological study by Huang et al. (2003) and Wallis et al. (2003). The coexisting biotite and muscovite in sillimanite-grade pelitic schist ($T > 600$ °C) give consistent cooling ages of ca. 40 Ma. On the other hand, biotite and muscovite in the lower-grade zones ($T < 500$ °C) yield discordant ages that are due to excess argon inherited from pre-metamorphic phases. They suggested that muscovite and biotite in polymetamorphic terranes require higher metamorphic temperatures to completely reset the Ar isotopic system than generally thought. Therefore, it should be noted that most rocks of the Tia Complex were formed at temperatures lower than the closure temperature of muscovite K–Ar system that is applicable to the phengite system. This requires the white micas in most metamorphic rocks show the ages of the high-P/T type metamorphism in the subduction zone and/or the timing of exhumation of the high-P/T type of rocks, at least in the rocks lower than the garnet-bearing zone. However, the white mica ages get younger gradually toward the granodiorite from 340 Ma to 260 Ma. This younging trend toward the granodiorite could be partly due to the grain size dependence of the closure temperature because of the very fine-grained micas in the metamorphosed accretionary complex. On the other hand, this could be also due to the chemical reaction of white micas by the thermal overprint of the high-P/T type metamorphic rocks and the associated argon depletion from mica crystal. Recrystallization with significant re-arrangement of major elements such Al and Si in the white mica crystal should involve argon release from white mica as the trapped argon is not favored in the K site of white micas, and easily diffuses out from white mica crystal structure. Fig. 2 shows that the lowest grade schist (No. 39; 339 Ma) has pure phengites and the highest-grade gneiss (No. 19; 259 Ma) has pure muscovite. The medium grade rocks in the higher A zone and the B zone have white micas with wide variation of Si content, suggesting that the chemical reaction of white micas proceeded in heterogeneous reaction domains with the thermal overprinting of the high-P/T type rocks. It should be noted there is gradual age change around the garnet-in boundary though the Si values have a drastic change.

6.2. Tectono-metamorphic evolution of the Tia Complex

As described above, the host lithologies of the low-P/T type of metamorphic rocks (ca. 260 Ma) in the Tia Complex were the high-P/T type of metamorphic rocks (ca. 340 Ma) formed in the subduction zone. This strongly suggests that the early Carboniferous subduction complex suffered the low-P/T type of metamorphism in Permian. This regional thermal overprinting of the subduction complex has been observed in the SW Japan consisting of the typical Pacific type orogenic belts. The Cretaceous Ryoke metamorphic sequences consist of the low-P/T type of metamorphic rocks of which the host lithologies are the Jurassic accretional complex (cf. Isozaki et al., 2010). They are associated with the peralminus granitic rocks that occur mainly in the higher-grade zone. Since Miyashiro (1961), Japanese geologists have believed that the sillimanite grade rocks of the Ryoke metamorphic belt occurred in the lower crust under the active volcanic arc where the arc magma rise up to heat the lower crust to produce its partial melting (cf. Isozaki et al., 2010). In the Tia Complex, Dirks et al. (1992) described a thermal overprint of a subduction–accretion complex in a compressional back-arc setting that occurred after the subduction zone shifted east in Permian. They did not take care of the Tia granodiorite that forms part of the Hillgrove Suite, which consists of S-type plutons (Flood and Shaw, 1977; Shaw and Flood, 1981).

We suggest that the S-type granitic rocks form by the partial melting of the subduction–accretion complex that suffer the low-P/T type metamorphism with the heating by a specific magma produced in the wedge mantle. Cretaceous high-Mg andesites and adakites have been observed in SW Japan (Ohira, 1995; Imaoka et al., 2011). This type of magma is distinct from the typical arc magma produced during the active subduction of oceanic plate. We consider the low-P/T type of Ryoke metamorphic rocks associated with the granitic rocks is due to the heating by the specific magma of the high-Mg andesite and adakites because the ages of the metamorphic rocks, granitic rocks and the volcanic rocks are consistent with each other. These specific magmas could be formed by the tectonic change in the wedge mantle such as the rollback or shift back of the subducting slab of oceanic plate and the subsequent upwell of high-T asthenosphere. In the NEFB, the specific volcanic rocks such as high-Mg andesite and adakite have not been observed yet. However, Waterhouse and Sivell (1987) and Murray (1990) described the east side shift of the subducting slab of oceanic plate in Permian. Recently, Rosenbaum et al. (2012) described U–Pb geochronology of Early Permian granitoids in the southern NEFB and proposed a tentative tectonic model that involves an early stage of subduction curvature during slab rollback at 300–285 Ma. These suggest the Early Permian tectonic change in the wedge mantle to produce the specific magma as seen in SW Japan. Fig. 3 shows a model for the tectono-metamorphic evolution of the Tia Complex. In Carboniferous, a steady state subduction of oceanic plate had taken place to form accretionary complex in eastern edge of Australia continent and coeval high-P/T metamorphic rocks in a deeper part of the complex. The east side slab rollback took place at 300–285 Ma, resulting in the subsequent upwell of high-T asthenosphere. This high-T asthenosphere underplated below the Carboniferous accretionary complex to form the specific magma such as high-Mg andesite and adakite, resulting in the thermal overprinting of accretionary complex and the partial melting of complex to form S-type granodiorites. We should look for the Permian specific volcanic rocks such as the high-Mg andesites and adakites in the New England Fold Belt in the future.

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