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METAMORPHISM AND GRANITE GENESIS IN THE HIDAKA METAMORPHIC BELT, HOKKAIDO, JAPAN

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Abstract: The Main Zone of the Hidaka Metamorphic Belt is an uplifted crustal section of island arc type. The crust was formed during early Tertiary time, as a result of collision between two arc-trench systems of Cretaceous age. The crustal metamorphic sequence is divided into four metamorphic zones (I to IV), in which zone IV is in the granulite facies.
A detailed study of the evolution of the Hidaka Belt, based on a revised P-T-t analysis of the metamorphic rocks, notably a newly found staurolite-bearing granulite, confirms a prograde isobaric heating path, after a supposed event of tectonic thickening of accretion sedimentary and oceanic crustal rocks. During the peak metamorphic event (ca. 52.6 Ma), the regional geothermal gradient attained 30 to 40° C/km, and the highest P, T condition obtained from the lowest part of the granulite unit is 830° C, 7kb. In this part, X_{H_2O} of Gt-Opx-Cd gneiss is about 0.15 and that of Gt-Cd-Bt gneiss is 0.4. The P-T-X_{H_2O} condition of the granulite unit is well within a field where fluid-present partial melting of pelitic and greywacke metamorphic rocks take place. This is in harmony with the restitic nature of the Gt-Opx-Cd gneiss in the lowest part of the granulite unit.
The possibility that partial melting took place in the Main Zone is significant for the genesis of the peraluminous (S-type) granitic rocks within it. The S-type granitic rocks in this zone are Opx-Gt-Bt tonalite in granulite zone, Gt-Cd-Bt tonalite in amphibolite zone, and Cd-Bt-Mus tonalite in the Bt-Mus gneiss zone. The mineralogical and chemical natures of these strongly peraluminous tonalitic rocks permit us to regard them to have been derived from S-type granitic magma generated by crustal anatexis of pelitic metamorphic rocks in deeper crust.

Key words: metamorphism, granite genesis, partial melting, P-T conditions, S-type granite, Hidaka Metamorphic Belt, Japan.

Introduction

In Hokkaido, north Japan, late Cretaceous accretionary complexes belonging to different arc-trench systems are juxtaposed in the central part of Hokkaido where the Hidaka Metamorphic Belt occurs (Fig. 1). A collision between the two arc-trench systems seems to have taken place in the early Tertiary resulting in the eastern accretion complex being thrust over the western complex (Komatsu et al., 1989).

The Hidaka Metamorphic Belt consists of two different zones: the Western Zone and the Main Zone (Komatsu et al., 1982, 1983). The former is composed of meta-ophiolitic rocks, which are faulted into slices and overturned (Miyashita, 1983; Jolivet & Miyashita, 1985). On the other hand, the Main Zone consists of various metamorphic rocks and acidic to basic intrusive rocks, and is regarded as a crustal slab of island-arc type (Komatsu et al., 1983). The boundary between the two zones is a large thrust fault (Hidaka Main Thrust: H. M. T.) which is associated with the mylonite zone over the whole extention of the Hidaka Metamorphic Belt.

The western side of the Hidaka Metamorphic Belt is bordeded by the Western Boundry Thrust (W. B. T.) and is bounded by the unmetamorphosed sediments of Idonnappu Group of the western Cretaceous accretional complex. On the eastern side of the belt, however, the metamorphic rocks of the Main Zone grade into the very low-grade to unmetamorphosed sediments of Nakanogawa Group of the eastern Cretaceous accretional complex.

Osanai (1985) investigated the detailed mineral parageneses of the metamorphic rocks in the central area of the Main Zone, and confirmed that the metamorphic grade increases progressively westward from greenschist up to granulite facies. Progressive changes in metamorphic grade shows the metamorphic field gradient (Spear et al., 1984) or the geothermal gradient to be 33–40 °C/km (Osanai et al., 1986; Komatsu et al., 1989). Komatsu et al. (1989) evaluated the estimated P-T conditions of metamorphism in relation to deformation events and igneous activities to establish the pressure-temperature-time-deformation (P-T-t-D) path of the Main Zone. Recently, staurolite break-down reactions,

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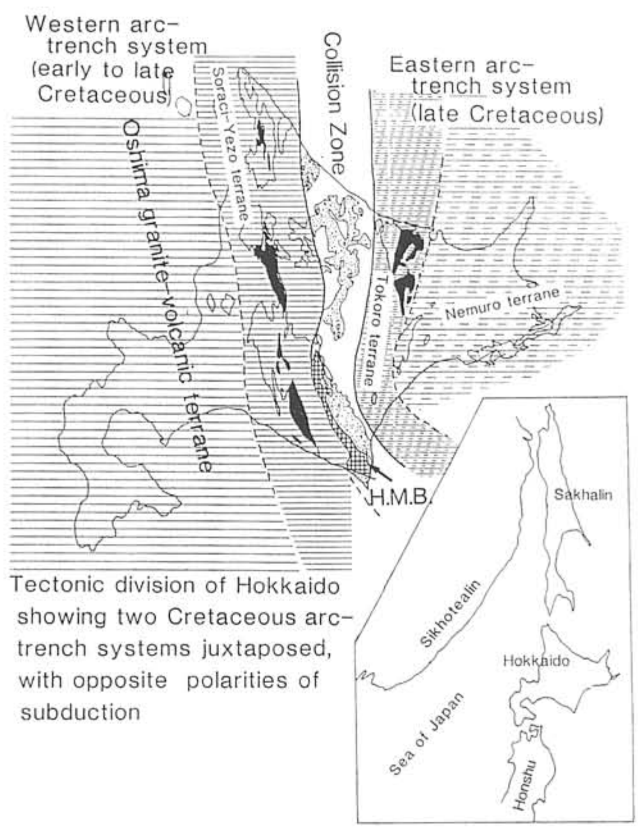

Fig. 1. Tectonic division of Hokkaido showing two Cretaceous arc-trench systems juxtaposed, forming a collision zone along the central part of Hokkaido (Komatsu et al., 1989). The southern and of the collision zone was evolved to the Hidaka Metamorphic Belt (H. M. B.) where island-arc type magmatism and metamorphism took place.

which show the prograde P-T path of the Main Zone, were found in pelitic rocks of granulite facies (Osanai, 1988). We combine these studies to shed a new light on the P-T-d-D path of the Main Zone with emphasis on the generation of S-type granitic intrusive rocks.

Geology of the Main Zone

The Main Zone, which is composed of various metamorphic and intrusive rocks, can be divided into two lithologic sequences: the lower metamorphic sequence (Lower sequence) rich in basic rocks and the upper metamorphic sequence (Upper sequence) consisting of pelitic and psammitic rocks (Komatsu et al., 1982). The Lower sequence is divided into granulite, amphibolite, hornblende-biotite gneiss units from west (lower structural level) to east (upper), and the Upper one into biotite-muscovite gneiss to schist and chlorite-muscovite metasedimentary rock units from west (lower) to east (upper) (Osanai, 1985). However, the boudary between the two sequences seems to have formed tectonically by thrusting of the latter over the former sequence before deformation and metamorphism because: (1) gneiss-foliations in both sequences are parallel, but those of the Lower sequence are also parallel to original lithological banding, while (2) those of the Upper sequence are oblique to it.

The zonal arrangement of metamorphic rocks from west to east throughout both sequences is thought to represent the lithologic sequence from lower to upper parts of island-arc type crust. A representative distribution of metamorphic and intrusive rocks in the Main Zone is shown in Fig. 2.

Upper metamorphic sequence

The biotite-muscovite gneiss and schist unit (ca. 5 km thick) consists of biotite-muscovite gneiss in the lower portion and biotite-muscovite-chlorite schist in the upper portion. Thin intercalations of cordierite-andalusite-biotite gneiss occur in the unit.

The muscovite-chlorite metasedimentary rock unit (ca. 4 km) is composed largely of muscovite- and chlorite-bearing, weakly foliated metasedimentary rocks. Schistose metasedimentary rocks including biotite, occur in the lower part of the unit. The original rocks of the Upper sequence may have been derived from sedimentary rocks of the Nakanogawa Group as evidenced by the gradational metamorphic change into the very low grades.

Lower metamorphic sequence

The granulite unit (0.5–1.5 km) consists of hypersthene amphibolite, brown hornblende amphibolite, hypersthene--plagioclase gneiss, garnet-hypersthene gneiss, garnet-hypersthene-cordierite gneiss, and garnet-cordierite-biotite gneiss.

The amphibolite unit (0.5–2.0 km) consists essentially of brown hornblende amphibolite with minor garnet amphibolite in the basal part of the unit. Some thin layers (10–30 m) of cordierite-gedrite gneiss and garnet-cordierite-biotite gneiss occur in the basal to middle part, and garnet-biotite gneiss occurs in the middle to upper part of the unit.

The hornblende-biotite gneiss unit (2–3 km) is composed largely of thin alternations of biotite gneiss and biotite amphibolite. Frequent intercalations of biotite gneiss, amphibolite and garnet-biotite gneiss are found in the basal part of the unit, while small amounts of biotite-muscovite gneiss are recognized in the upper part of the unit.

S-type granitic rock intrusions

The intrusive bodies of S-type granitic rocks in the Main Zone can be classified into three types in terms of their characteristic features such as differences in emplacement horizon, lithofacies and the kind of their inclusions, as follows (Komatsu et al., 1986):

1. Biotite-muscovite and biotite-muscovite-cordierite heterogeneous tonalite (middle tonalite) which intruded as large masses between the Lower and Upper sequences, and is characteristically surrounded by layered-type migmatite. The inclusions in this tonalite are biotite-muscovite gneiss and schist, garnet-biotite gneiss and amphibolite. Sometimes, they are accompanied by homogeneous hornblende-bearing tonalite.
2. Garnet-cordierite heterogeneous tonalite (lower tonalite) which occurs in the amphibolite unit of the Lower sequence, and contains inclusions of orthopyroxene-bearing pelitic and basic granulites and brown hornblende amphibolite.
3. Hypersthene-garnet heterogeneous tonalite (basal tonalite) which is found in the granulite unit of the Lower sequence. Especially in the upper part of the granulite unit, it forms agmatite with pelitic and basic granulites. Inclusions are

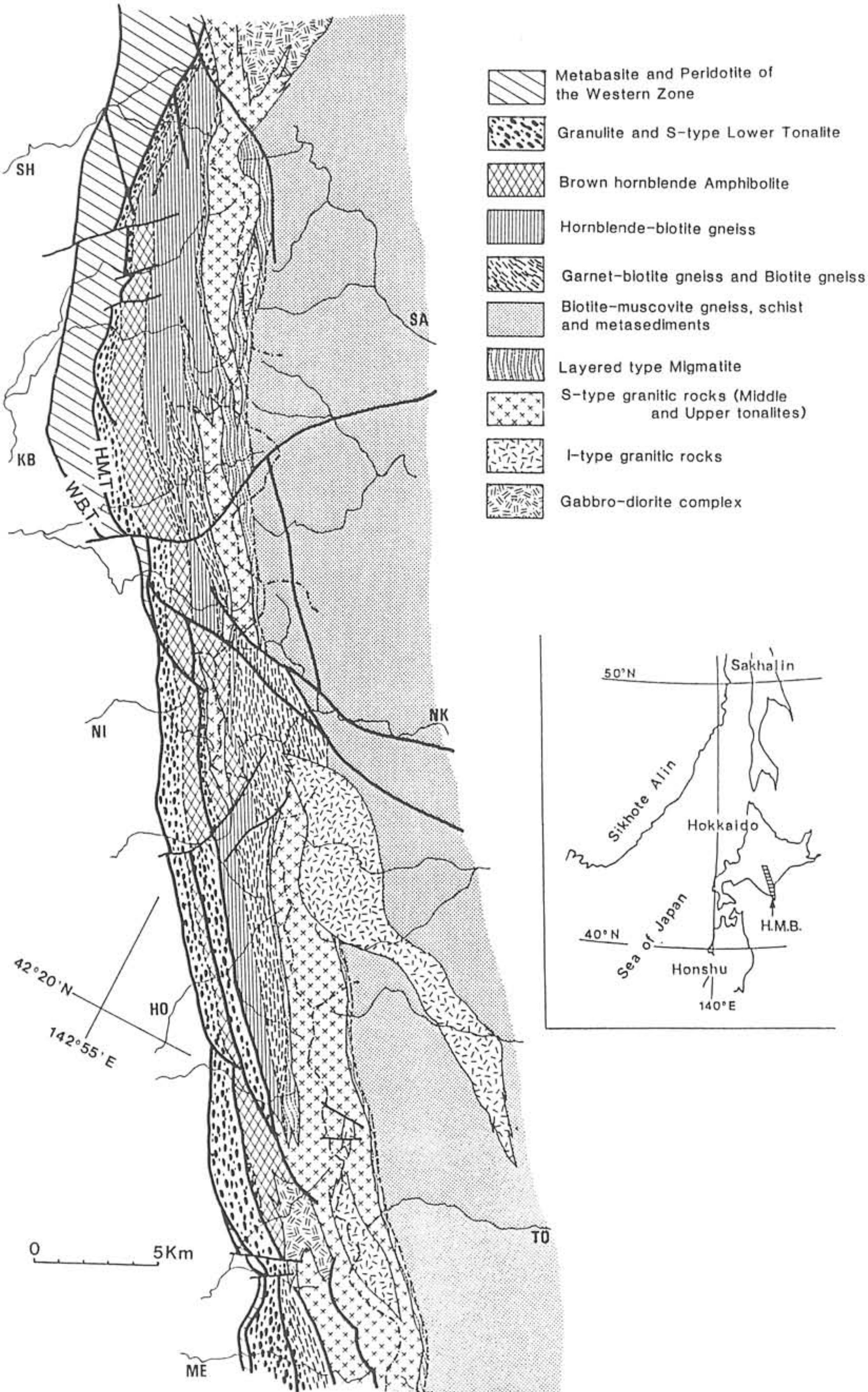

Fig. 2. Geologic map of the central area of the Hidaka Metamorphic Belt.
W.B.T.: Western Boundary Thrust, H.M.T.: Hidaka Main Thrust, SH: Shunbetsu river, KB: Koibokushu-Shibichari r., NI: Nishoumanai r., HO: Horobetsu r., ME: Menashunbetsu r., SA: Satsunai r., NA: Nakanogawa r. and TO: Toyoni r.

garnet-hypersthene gneiss, hypersthene-plagioclase gneiss, hypersthene amphibolite, brown hornblende amphibolite, and meta-harzburgite. The granitic rocks along the thrust fault of H. M. T. are remarkably mylonitized.

Metamorphism and deformation events of the Main Zone

Deformation process of the Main Zone

The metamorphic rocks of the Main Zone of the Hidaka Metamorphic Belt were affected by shearing and underwent varying degrees of retrograde metamorphism. Two later-stage deformational events (D1 and D2) can be defined based on geometrical nature of mylonites and petrologic nature of retrograde metamorphism (M1 and M2) of the deformed rocks. The early-stage deformation event (D0) is speculated as that of tectonic thickning of sedimentary rocks followed by igneous activity and progressive metamorphism (M0) to form an island-arc type crust during early Tertiary time as a result of collision between two arc-trench systems of Cretaceous age (Komatsu et al., 1989).

a) Subhorizontal displacement (D1)

Several layer-parallel shear zones occur at different levels of the metamorphic sequence and intrusives. Recrystallized mineral assemblages formed by the retrograde metamorphism (M1) through the deformation in these shear zones vary with the structural depth. They are in the granulite facies in the lower, amphibolite facies in the middle and lower-amphibolite to greenschist facies in the middle to upper portions of metamorphic sequence in the Main Zone. Recrystallized mineral lineations in any one shear zone are subparallel to those in other shear zones, and the sense of movement is dextral. This evidence indicates that shearing took place at different crustal levels due to subhorizontal displacement or low-angle thrust movement of crustal layers. During this movement, the lower to upper crustal layers may have been detached from the lowermost crustal layer.

b) Dextral-reverse movement (D2)

Mylonites are widely developed along the basal part of the Lower sequence and are in contact with the Western Zone of the Hidaka Metamorphic Belt. The mylonites form mostly in the basal heterogeneous tonalites. Intensity of mylonitization increases towards the H. M. T., accompanied by retrograde metamorphism (M2) down to greenschist facies condition. Mylonite foliations trend nearly parallel to the foliation of the metamorphic rocks and mineral lineations plunge very gently northwards. Minor structures indicate a dextral-reverse sense of movement (Arita et al., 1986). The structures in the Western Zone are parallel to those of the basal mylonites in the Main Zone and show the same sense of movement (Jolivet and Miyashita, 1985).

Regional metamorphic zonation

The Main Zone can be divided into four metamorphic zones from the top (east) to the base (west) in terms of the mineral parageneses of pelitic and psammitic rocks formed during the progressive metamorphic event (M0). The Upper sequence is assigned to zone I and the low-grade part of zone II, while the Lower sequence refers to the high-grade part of zone II, zone III, and zone IV. Zones III and IV show a notable difference in mineral parageneses between pelitic and psammitic rocks. Gt-Opx-Cd gneiss in zone IV is possibly a restite left after partial melting of pelitic gneiss. In the Lower sequence, basic rocks also change in mineral parageneses consistent with paragenetic change in pelitic rocks. The critical mineral parageneses are as follows (Osanai et al., 1986; Komatsu et al., 1989):

for pelitic rocks
  zone I: Chl, Phe
  II: Bt, Mus, (And, Gt or Cd)
  III: Bt, Gt, Sil, (Cd or Kfs)
  IV: Gt, Cd, Kfs, (Sil or Bt)
for psammitic rocks
  III: Bt, Gt, Cd, Ged
  IV: Gt, Cd, Opx, Bt
for basic rocks
  II: green brown Hbl, Bt, Ep
  III: brown Hbl, Cum, (Bt, Gt)
  IV: brown Hbl, Cpx, Opx

The regional distribution of each zone in the central area of the Main Zone is shown in Fig. 3. These progressive changes in metamorphic grade show a metamorphic field gradient (Spear et al., 1984) or an island-arc type geothermal gradient of 33–40 °C/km (Osanai et al., 1986; Komatsu et al., 1989).

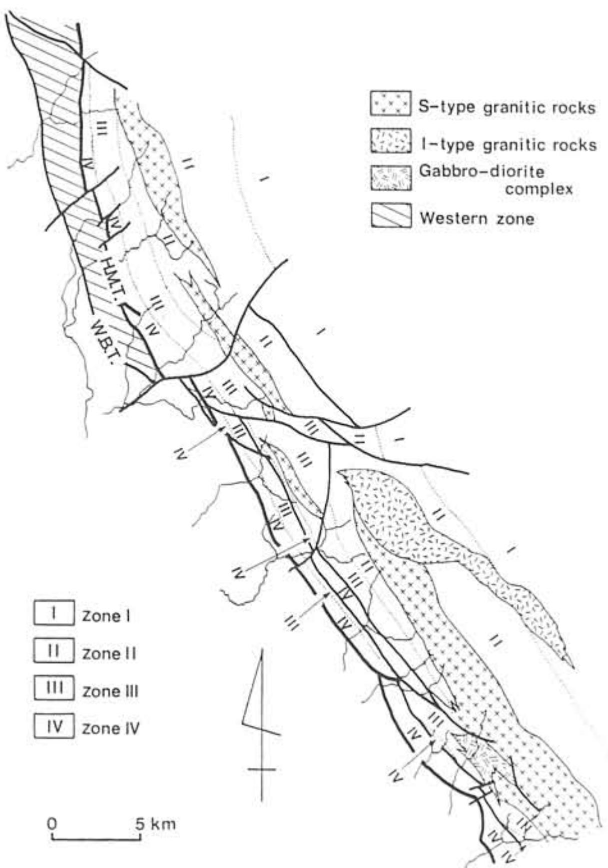

Fig. 3. Regional metamorphic zonation of the central area in the Main Zone. The area is the same as in Fig. 2.

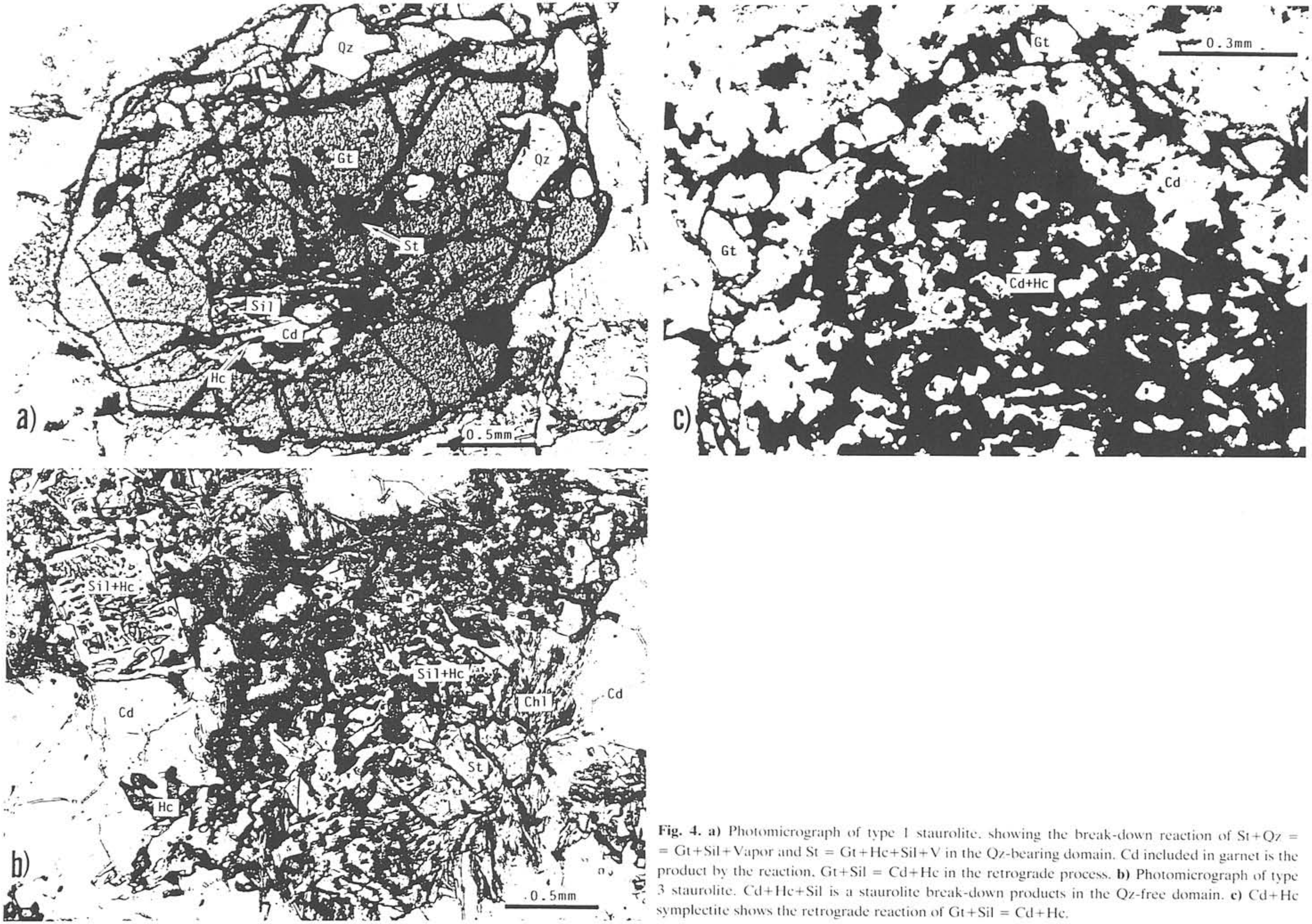

Fig. 4. a) Photomicrograph of type 1 staurolite, showing the break-down reaction of St+Qz = = Gt+Sil+Vapor and St = Gt+Hc+Sil+V in the Qz-bearing domain. Cd included in garnet is the product by the reaction. Gt+Sil = Cd+Hc in the retrograde process. **b)** Photomicrograph of type 3 staurolite. Cd+Hc+Sil is a staurolite break-down products in the Qz-free domain. **c)** Cd+Hc symplectite shows the retrograde reaction of Gt+Sil = Cd+Hc.

Regional prograde metamorphism

The progressive metamorphism (M0) along the estimated geothermal gradient indicates the maximum metamorphic conditions of each zone during the D0 event after island-arc type magmatic activity (Komatsu et al., 1989). The prograde metamorphic process before M0 can be evaluated only in the granulite unit of zone IV. In granulite-facies rocks the chemical compositions of minerals are generally homogeneous, except for marginal chemical zoning of minerals caused by retrograde metamorphism. Therefore, the prograde metamorphic process is usually analysed by using phase relations among disequilibrium minerals found as relict inclusions.

Osanai (1988) reported the presence of three-types of staurolite in the Main Zone, and these should be valuable in investigating the prograde metamorphic process.

The staurolite-bearing garnet-cordierite-sillimanite gneiss occurs as a layer about 10 m thick enclosed in hypersthene amphibolite of zone IV. There are three modes of occurrences of staurolite; (1) type 1 (Mg/Fe + Mg + Zn: X_{Mg} = 0.25−0.33, Zn/Fe + Mg + Zn: X_{Zn} = 0.02−0.03): inclusions in pyrope-almandine garnet (Fig. 4); (2) type 2 (X_{Mg} = 0.21−0.26, X_{Zn} = 0.03−0.04): inclusions in plagioclase; and (3) type 3 (X_{Mg} = 0.16−0.20, X_{Zn} = 0.04−0.10): porphyroblasts associated with cordierite, hercynite, corundum and sillimanite (Fig. 4). Staurolite in all three modes can be observed within one thin section. However, the rock is subdivided into two domains of quartz-bearing and quartz-free domains. Types 1 and 2 staurolite occur in the former domain, while type 3 occurs in the latter one.

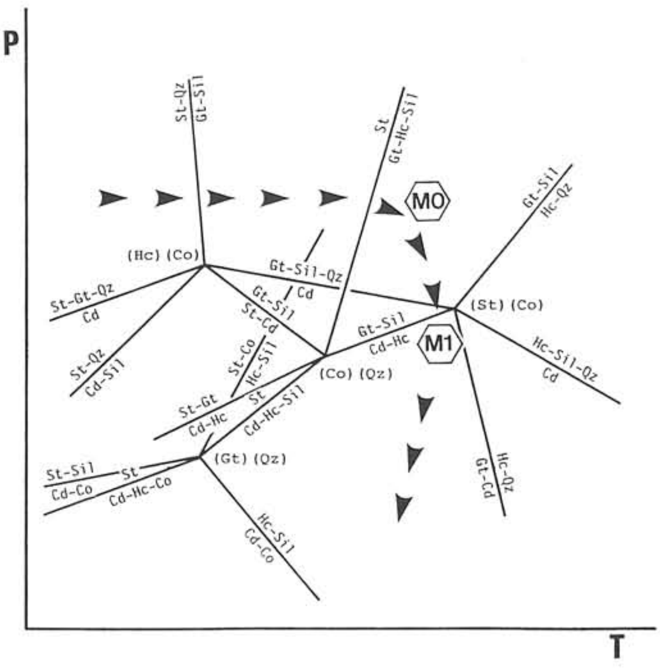

Fig. 5. Schematic phase relations of staurolite and related minerals in part of $FeO-Al_2O_3-SiO_2-H_2O$ system.
Arrows indicate possible prograde and retrograde path of the quartz-free domain in pelitic granulite of zone IV. As MgO and Zn are not taken into account in this system, the network may shift to higher P-T side.

In the quartz-bearing domain, we can observe the staurolite (type 1) break-down products of garnet-sillimanite and garnet-sillimanite-hercynite caused by prograde metamorphic reactions (Fig. 5). Furthermore, in the quartz-free domain, the prograde reaction including staurolite (type 3) proceeds to form hercynite-sillimanite and cordierite-hercynite-sillimanite under the same P-T change as type-1 staurolite. These prograde reactions had taken place during nearly isobaric heating before M0 metamorphism (Osanai, in prep.).

Contact metamorphism in the southern area of the Main Zone

Owada (1989) showed that the upper metamorphic rocks of zone II were affected by contact metamorphism up to low-pressure granulite facies of 700 °C and 4kb caused by intrusion of S-type tonalitic rocks in the southern end of the Main Zone. The contact aureole, which is named the sillimanite zone in Owada (1989), and is equivalent to zone D of Shiba (1988), is ca. 100 m wide, surrounding S-type granitic rock masses, where partial melting of pelitic and basic rocks took place (see Tagiri et al., in this issue).

Retrograde metamorphism in the Main Zone

The retrograde metamorphism is recognized mainly in the granulite-facies rocks of zone IV. In the pelitic granulite, the rims of garnet and cordierite are replaced by fibrous biotite and sometimes by muscovite and chlorite. The garnet-sillimanite association in staurolite-bearing gneiss was replaced by hercynite-cordierite symplectite, which is also a retrograde assemblage (Fig. 4). In the psammitic granulite, garnet and orthopyroxene are replaced by biotite, muscovite and chlorite, and by anthophyllite or gedrite, respectively. Furthermore, in the calc-silicate rocks in zone IV, wollastonite was broken down to grossular-quartz, followed by prehnite and pumpellyite formed at the minimum condititon of ca. 300 °C and ca. 2 kb. The retrograde processes are divided into two stages; e. g. 1) decompression with slight cooling (M1 from D0 to D2), and 2) nearly isobaric cooling (M2 after D2) (Fig. 7).

Metamorphic P-T conditions

P-T conditions of greenschist-amphibolite facies transitional zone

This zone is equivalent to the transitional zone between zones I and II.

According to Miyoshi (1986), d(002) and Lc(002) values of carbonaceous material change drastically between zones I and II. In zone III, however, the graphite is fully ordered. In comparison with the experimental data of Landis (1971) and Tagiri and Oba (1983), the metamorphic temperatures of the transitions between zones I and II, and zones II and III are ca. 400 °C and ca. 500 °C, respectively.

The biotite-muscovite-chlorite-quartz geobarometer by Powell and Evans (1983) may be employed for the pressure estimation of zone II, where the distribution coefficient between biotite and chlorite is 0.84 and gave 2.6−3.5 kb.

Table 1. P-T estimates of amphibolite-facies (zone III) rocks

Temperature

Sample No.	X_{Fe}^{Gt}	X_{Mg}^{Gt}	X_{Ca}^{Gt}	X_{Mg}^{Bt}	X_{mg}^{Hb}	T1	T2	T3
Upper part								
73021	0.604	0.166	0.027	0.561	—	538	551	—
73025	0.621	0.142	0.031	0.526	—	578	583	—
Lower part								
80616	0.652	0.167	0.028	0.519	—	622	616	—
61104	0.688	0.220	0.038	0.557	—	709	645	—
8418	0.703	0.198	0.093	—	0.424	—	—	664
BKGA1	0.709	0.207	0.074	—	0.420	—	—	641

Pressure

Sample No.	X_{Mg}^{Gt}	X_{Ca}^{Gt}	X_{Mg}^{Cd}	X_{Ca}^{Pl}	P1(Fe)	P1(Mg)	P1(Ave)	P2	T*
73025	0.142	0.031	0.624	—	5.3	5.0	5.1	—	580
80616	0.167	0.028	0.620	—	5.2	5.9	5.5	—	620
61104	0.220	0.038	—	0.388	—	—	—	5.6	670

Methods for temperature and pressure estimation are as follows; T1: Thompson (1976), T2: Perchuk (1977), T3: Graham and Powell (1984), P1: Wells (1979), P2: Newton and Haselton (1981). T* is mean temperature of each sample.

Table 2. P-T estimates of granulite-facies (zone IV) rocks

Temperature (P=6.5 kb)

Sample No.	X_{Mg}^{Gt}	X_{Ca}^{Gt}	X_{Mg}^{Bt}	X_{Mg}^{Cd}	X_{Mg}^{Opx}	T1	T2	T3	T4	T5
72302	0.272	0.020	0.591	0.680	—	740	670	625	813	—
SBT81	0.261	0.020	0.530	0.698	—	811	721	764	756	—
P8415-2	0.296	0.026	0.564	0.719	—	839	735	764	787	—
81003	0.311	0.058	0.612	—	0.533	803	714	—	—	770
7806	0.328	0.030	0.632	0.743	0.578	789	700	779	801	750
SS31(SS11)	0.293	0.18	0.547	0.688	—	849	748	807	841	—

Pressure

Sample No.	X_{Mg}^{Gt}	X_{Ca}^{Gt}	X_{Mg}^{Opx}	P1	P2	
72302	0.272	0.020	0.171	—	7.1(750)	—
SBT81	0.261	0.020	0.215	—	6.5(800)	—
P8415-2	0.296	0.026	0.238	—	7.3(800)	—
81003	0.311	0.058	0.384	0.537	—	6.9(800)
7806	0.328	0.030	0.228	0.578	—	6.1(780)
SS11(SS31)	0.352	0.019	0.167	0.580	—	7.2(800)

Methods for temperature and pressure estimation are as follows; T1: Thompson (1976), T2: Perchuk (1977), T3: Thompson (1976), T4: Perchuk et al. (1983), T5: Harley (1984), P1: Newton and Haselton (1981), P2: Newton and Perkins (1982).

P-T conditions of amphibolite facies metamorphism

The amphibolite facies zone is mainly equivalent to zone III. For the estimation of metamorphic temperature conditions, we employed the following Fe-Mg exchange geothermometers; 1) garnet-biotite (Thompson, 1976; Perchuk, 1977) and 2) garnet-hornblende (Graham and Powell, 1984). The former is used for pelitic metamorphic rocks, the latter for basic rocks such as garnet amphibolite.

Temperature calculations based on Thompson's and Perchuk's methods show nearly the same values. The temperatures of structurally upper part of zone III are 530−580 °C, while the temperature of lower part increases up to 670 °C.

The estimated temperatures obtained by Graham and Powell's method are 600−670 °C which are nearly consistent with the garnet-biotite temperatures.

For estimation of the peak metamorphic pressure conditions, the following two geobarometers can be employed for the intercalating garnet-cordierite-sillimanite-biotite gneiss; 1) garnet-plagioclase-Al_2SiO_5-quartz (Newton and Haselton, 1981; Ganguly and Saxena, 1984), and 2) garnet-cordierite-Al_2SiO_5-quartz (Wells, 1979).

The pressure conditions are estimated up to be 5.6 kb by the former method, but 5.2−6.0 kb using the Fe-endmember reaction and 5.0−5.9 kb for Mg-endmember (5.1−5.5 kb in average) by the latter method. The results are listed in Tab. 1.

Table 3. Temperature estimates of basic granulite

Sample No.	a_{Mg}^{Cpx}	a_{Mg}^{Opx}	X_{Fe}^{Opx}	T1	T2
71508	0.032	0.286	0.449	806	799
NNC8	0.040	0.334	0.410	836	828
SP130	0.032	0.167	0.578	873	810

T1: Wood and Banno (1973), T2: Wells (1977).

P-T conditions of granulite facies metamorphism

The granulite-facies zone is equivalent to zone IV. Temperatures have been estimated using Fe-Mg exchange between garnet-biotite (same as before and the method of Indares and Martignole, 1985), garnet-cordierite (Perchuk et al., 1983; Thompson, 1976; Holdaway and Lee, 1977) for pelitic rocks, garnet-orthopyroxene (Harley, 1984) and garnet-cordierite for psammitic rocks and 2-pyroxene (Wood and Banno, 1973; Wells, 1977) for basic rocks.

The Perchuk thermometer gives lower temperatures, while average values of Thompson's and Indares and Martignole's methods seem reasonable (740−840 °C). The garnet-cordierite (above mentioned) and garnet-orthopyroxene (Harley, 1984) thermometers give nearly consistent with or slightly lower than the garnet-biotite thermometry (720−840 °C). 2-pyroxene thermometries give slightly higher temperature (790−870 °C) than the others. Pressures were estimated using the garnet-plagioclase-sillimanite-quartz barometer (Newton and Haselton, 1981) for pelitic rocks (5.7−7.3 kb), and the garnet-orthopyroxene-plagioclase--quartz barometer of Newton and Perkins (1982) for psammitic rocks (6.1−7.2 kb). The results are listed in Tabs. 2 and 3.

The P-T conditions of retrograde metamorphism were estimated using the garnet-biotite thermometer and garnet-

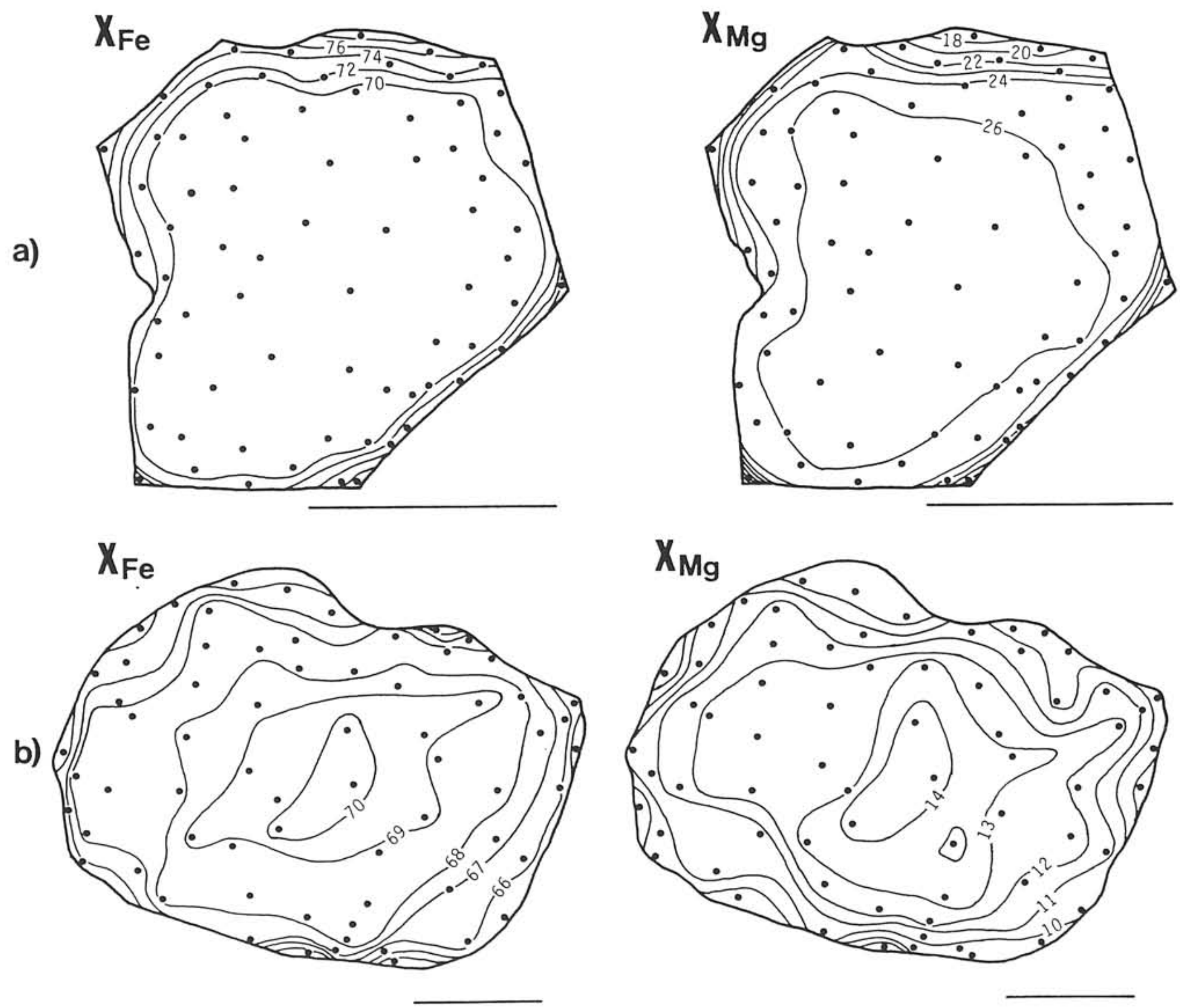

Fig. 6. Composition map of garnet in pelitic granulite. **a)** Garnet from undeformed rock; **b)** garnet from deformed rock. Dots indicate analysed points. Numbers in figures show almandine and pyrope molecule, respectively. Scale bars indicate 100 μm.

-cordierite-spinel-quartz barometer (Bhattacharya and Sen, 1986) for pelitic rocks of zone IV.

Almost all the Fe-Mg minerals are chemically zoned, especially garnet which shows strong marginal zoning when in contact with other Fe-Mg minerals. For undeformed rocks, core compositions of minerals often preserve higher conditions than rim compositions, whereas for deformed rocks core compositions give similar P-T values to those of the rim of undeformed minerals (Fig. 6). Furthermore, as both core and rim compositions vary grain by grain in the same section, estimated P-T values range very widely, for example, from 560 to 850 °C and from 3.0 to 7.2 kb. We regarded the highest grade core P-T sets from undeformed rocks to represent the conditions of an early event (M0), and rim P-T sets from the undeformed rocks and core P-T sets from deformed rocks as conditions of the second event (M1). Rim P-T sets from deformed rocks are grouped as conditions of the third event (M2).

P-T-t-D part of the Main Zone

Komatsu et al. (1989) distinguished five stages of evolution of the Hidaka Metamorphic Belt: (1) tectonic thickening of sedimentary rocks of Cretaceous accretionary complex to form a sedimentary pile as a result of collision between two arc-trench systems, (2) island-arc type magmatic activity followed by granulite-facies metamorphism under a geothermal gradient of ca. 40 °C/km (M0 metamorphic event). This high gradient was the result of early stage gabbro-diorite and granodioritic magmatism, (3) subhorizontal movement (D1) accompanied by shearing and retrograde metamorphism (M1), (4) dextral-reverse movement (D2) accompanied with shearing and retrograde metamorphism (M2), followed by (5) uplift and mountain building. The P-T-t-D path of the Main Zone is shown in Fig. 7.

At the first stage, metamorphic pressure condition of the lower part of the sedimentary pile would be high, while temperature was still low. Subsequently, isobaric heating with prograde metamorphism up to granulite facies conditions (MO) was caused by magmatism between stages 1 and 2. The age of this magmatism is 52.6 ± 6.8 Ma by the Rb-Sr whole rock isochron dating (Owada and Osanai, 1988). In the third stage, the crustal layer of the southern area of the Main Zone formed a nappe at the age of 40.3 ± 0.8 Ma by Rb-Sr mineral isochron method (Owada and Osanai, 1988). However, in the central area of the Main Zone, it is assumed that the nearly isothermal decompression took place at the same age. K-Ar mineral ages were obtained for hornblende (17.5 ± 0.9 Ma), biotite (16.3 ± 0.6 – 17.1 ± 0.6 Ma) and muscovite (17.1 ± 0.4 Ma) from the basal mylonite and hornblende-biotite gneiss of the Main Zone and pelitic schist of the Western Zone (Shibata et al., 1984). These ages indicate the cooling age of each mineral in the fourth stage. Another cooling age (19.9 ± 0.4 Ma) was obtained by the Rb-Sr whole rock-mineral ischron dating for pelitic gneiss of zone IV (Osanai, unpub.). The relation between these ages and closure temperatures were plotted on the retrograde P-T path of Fig. 7. The last stage of the evolution, conglomerate deposition, which was derived from the Hidaka Metamorphic Belt caused by the mountain building, was formed at Late Miocene time (Miyasaka, 1987). The detailes of the evolutional process along the P-T-t-D path is given by Komatsu et al. (1989).

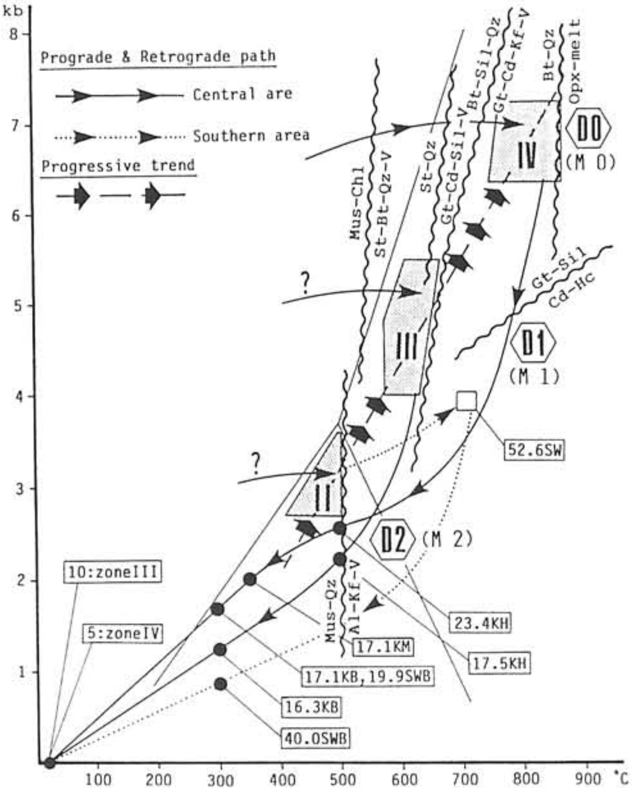

Fig. 7. P-T-t-D path of the Main Zone of the Hidaka Metamorphic Belt. M0, M1 and M2 show the metamorphic events. Deformation events of D0, D1 and D2 are from Komatsu et al. (1989).
An arrowed dotted line shows the reheating process by granitic intrusion in the southern area. Al_2SiO_5 triple point is after Holdaway (1971). Al: Al_2SiO_5 mineral, Kf: K-feldspar, V: vapor. Other mineral abbreviations are shown in appendix. Abbreviations for radiometric dating:
K: A-Ar method, S: Rb-Sr method, W: whole rock isochlon age, H: hornblende mineral age, M: muscovite mineral age, B: biotite mineral age, and WB: whole rock-biotite isochron age. The excavation age of zone III and IV at the surface are determined by sedimentary age of conglomerate which include metamorphic rocks, after Miyasaka (1987).

Origin of S-type granitic rocks

Mineralogical and chemical character of granitic rocks

Three types of tonalitic rocks (middle, lower and upper) which intrude into the metamorphic sequences in the Main Zone contain, besides biotite, one or two peraluminous minerals: muscovite and/or cordierite in the middle, garnet and/or cordierite in the lower and garnet in the basal tonalite. Orthopyroxene is commonly present in the lower tonalite, and is one of main constituents in the basal tonalite. Therefore, these tonalites are a strongly peraluminous suite according to Miller's criteria (Miller, 1985). Most of the AFM minerals contained in these tonalites have, according to criteria by Wall et al. (1987), clear evidence of magmatic origin, but some grains of the same minerals exhibit textural and chemical features showing that they have been formed by the reaction between the magma and metamorphic enclaves, or were derived from country metamorphic rocks due to either mechanical breakup and mixing, or partial melting due to the intrusion of the granitic magma.

All these tonalites show S-type chemical affinities (White and Chappel, 1977). SiO_2 contents range from 59−70 wt % in the middle tonalite, 59−67 wt % in the lower tonalite and 57−68 wt % in the basal tonalite. In general, the lower tonalite is chemically very similar to the basal tonalite, and hereafter both are considered as a single entity for which the name, basal tonalite will be used. The middle tonalite is richer in K_2O and Al_2O_3 and poorer in MgO, FeO and CaO than the basal tonalite. The difference of these oxide contents is large compared with the difference of SiO_2 · $Al_2O_3/(CaO + Na_2O + K_2O)$ molar ratios are 1.0−1.5 for the middle tonalite and 0.95−1.25 for the basal tonalite. These chemical variations between the middle and basal tonalites are likely ascribed to fractional crystallization of a peraluminous granitic magma. This is supported by initial ratios of Rb-Sr isotopes in both tonalites which are in the same range (0.705−0.706) (Fig. 8). These initial ratios show nearly the same in those of pelitic metamorphic rocks (Fig. 8).

tonalitic magma, crystallized at high temperature (> 900 °C) and at moderate pressure (5−6 kb), and that the middle tonalite has been derived from the same magma emplaced at a higher level of the crust, modified by fractionation at depth and by assimilation of biotite-muscovite gneiss, crystallized at relatively low temperature and high f_{H_2O} and $a_{KAlSi_3O_8}$.

Anatexis of pelitic rocks in the granulite facies

P-T-X_{H_2O} relations

The quantitative estimation of X_{H_2O} can be made by using of stability of hydrous minerals. In this study, the estimation was made by using of biotite stability, following Ferry (1981) for biotite-K-feldspar-ilmenite-pyrrohtite assemblage of Gt-Cd--Bt gneiss and Bhattacharya and Sen (1986) for the orthopyroxene-biotite-K-feldspar-quartz subassemblage of Gt-Opx-Cd gneiss in zone IV. Furthermore, muscovite stability (Tyler and Ashworth, 1982) was used for the muscovite-biotite gneiss of zone III. The thermochemical

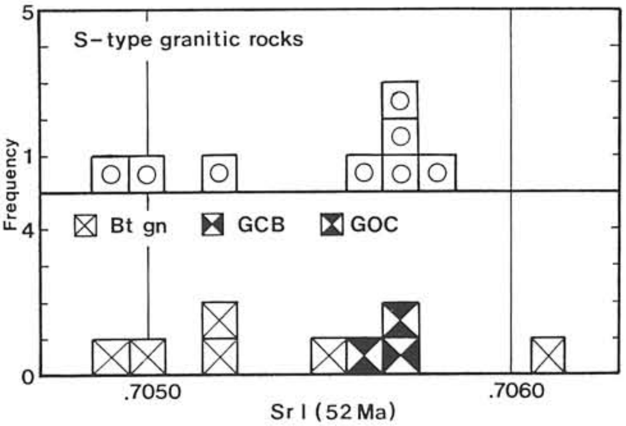

Fig. 8. ^{86}Sr/^{87}Sr initial ratios of S-type granites and metamorphic rocks.
Bt gn: Bt gneiss in the Upper Sequence, GCB: Gt-Cd-Bt gneiss in the Lower Sequence, GOC: Gt-Opx-Cd gneiss in the Lower Sequence. The initial ratios were calculated supporting the age of 52Ma (Owada and Osanai, 1988).

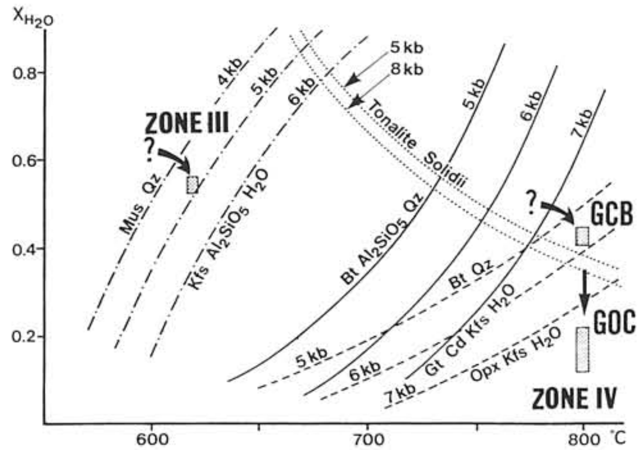

Fig. 9. T-H_{H_2O} relations for pelitic Gt-Bt gneisses of zone III and zone IV, and Gt-Opx-Cd gneiss in zone IV. Note that X_{H_2O} abruptly decrease from Gt-Cd-Bt gneiss to Gt-Opx-Cd gneiss in zone IV.

In contrast to the interpretation of Miller and Stoddard (1981) on the origin of magmatic garnet in peraluminous granite, garnet is predominant in the basal tonalite which is relatively less aluminous than the middle tonalite where cordierite is common. Textural relationships between minerals in the basal tonalite show that the order of crystallization from the magma was garnet-plagioclase-orthopyroxene-biotite and quartz. The observations are consistent with the experiment and interpretation by Clemens and Wall (1981, 1988) on the crystallization of peraluminous granitic magma, showing that garnet crystallizes at higher T and lower f_{H_2O} than cordierite, and that at high P and T, garnet crystallizes first from a peraluminous granitic magma. The early crystallization of garnet and plagioclase in the basal tonalite prior to orthopyroxene indicates relatively low a_{SiO_2} as well as low $a_{KAlSi_3O_8}$ and f_{H_2O} at the beginning of crystallization of the magma which formed the basal tonalite. It is concluded that the basal tonalite has been derived from peraluminous

parameters for the first method such as a_{FeS} and f_{S_2} are obtained from Toulmin and Barton (1964) and f_{H_2S} and f_{O_2} are obtained from Whitney (1984). Those for the second method such as Gibbs free energy and fugacity of H_2O are obtained from Robie et al. (1979) and Burnham et al. (1969), respectively. To calculate X_{H_2O}, we also use the fugacity coefficient of H_2O by Wood and Fraser (1979) and average P-T conditions as described above. The estimated values of X_{H_2O} are 0.53−0.57 for muscovite-biotite gneiss, 0.40−0.44 for Gt-Cd-Bt gneiss and 0.11−0.21 for Gt-Opx-Cd gneiss.

Fig. 9 shows the P-T-X_{H_2O} relations of some reaction curves for pelitic rocks in the Lower sequence, minimum melting curves and tonalite solidii. The Gt-Cd-Bt gneiss has chemical composition very similar to tonalite and the X_{H_2O} values of the gneiss are plotted at the high temperature side of the tonalite solidii obtained by Eggler (1972). Because of this, it is possible that in situ partial melting of the Gt-Cd-Bt gneiss took place at the condition of supposed P-T-X_{H_2O}.

Anatexis of pelitic metamorphic rocks

The highest metamorphic condition estimated is well within a field where anatexis under fluid present condition takes place in the pelitic metamorphic rocks of the Main Zone. Gt-Opx-Cd gneiss which occurs in the lowest part of the granulite unit presents evidence of anatexis. In general, Gt-Opx-Cd and Gt-Cd-Bt gneisses, which have nearly the same values of Sr initial ratios (Fig. 8), vary widely in chemical composition, but are in the same ranges of chemical composition as Bt gneiss in the upper metamorphic sequence and the uppermost, very low-grade, metasedimentary rocks of the Nakanogawa Group (Fig. 10). The Gt-Opx-Cd gneiss in the lowest part of the granulite unit, however, is richer in MgO and FeO and poorer in SiO_2 and especially in K_2O than

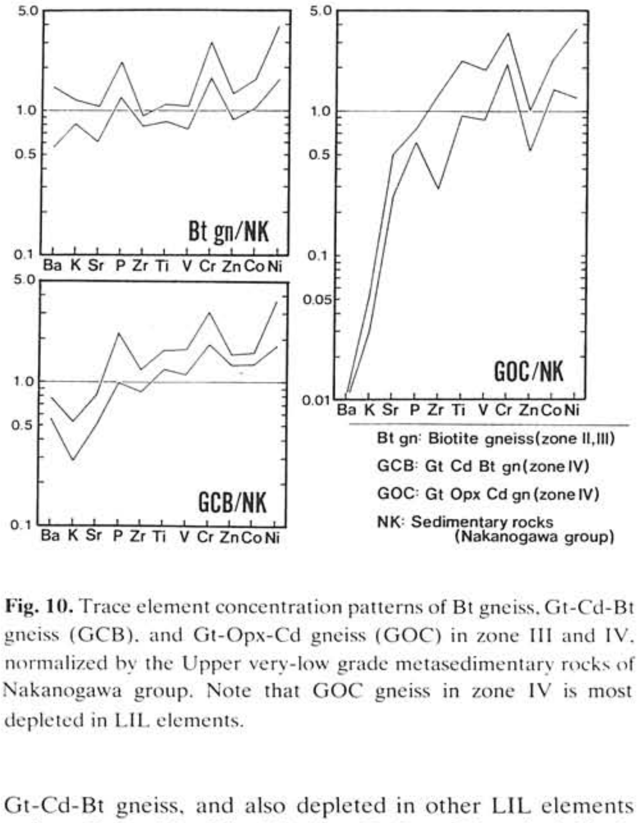

Fig. 10. Trace element concentration patterns of Bt gneiss, Gt-Cd-Bt gneiss (GCB), and Gt-Opx-Cd gneiss (GOC) in zone III and IV, normalized by the Upper very-low grade metasedimentary rocks of Nakanogawa group. Note that GOC gneiss in zone IV is most depleted in LIL elements.

Gt-Cd-Bt gneiss, and also depleted in other LIL elements such as Ba and Sr (Fig. 10). The Gt-Opx-Cd gneiss is likely a restite remained after partial melting of Gt-Cd-Bt gneiss. The reaction leading partial melting of Gt-Cd-Bt gneiss is,

$$\text{Bt} + \text{Qz} + \text{vapor} = \text{Opx} + \text{liquid}$$

$$\text{or Bt} + \text{Gt} + \text{Qz} + \text{vapor} = \text{Opx} + \text{liquid}$$

(Vielzeuf and Holloway, 1988). Kfs is not generally recognised or a very few amount, if any, in Gt-Opx-Cd gneiss. Applying the fluid-present reaction curve of Vielzeuf and Holloway (1988), the above reaction in the normal pelitic and greywacke with $X_{Mg} = 0.5$ takes place at around 830 °C, if we apply 7 kb as the pressure condition of the Gt-Opx-Cd gneiss. This temperature is very close to the highest one estimated for the Gt-Opx-Cd gneiss in the lowest part of granulite unit. At this condition, however, only incipient anatectic melting would be expected.

On the other hand, strongly peraluminous (S-type) tonalitic rocks occur, as stated earlier, forming large intrusive masses within lowest to middle metamorphic layers. The inferred temperature of peraluminous tonalitic magma is higher than the highest T condition of the granulite where the basal tonalite is emplaced. This means that the peraluminous tonalitic magma might have been generated in a further deep part of the crust. The evidence for incipient anatexis of pelitic gneiss in the granulite unit strongly suggests the possibility that peraluminous granitic magma might have been generated through anatexis of pelitic rocks in deeper crust.

Conclusions

A detailed study on the evolution of the Hidaka Metamorphic Belt based on a revised P-T-t analysis of metamorphic rocks from the Main Zone confirmed a prograde isobaric heating path, after supposed tectonic thickening of accreted terrigeneous sedimentary and oceanic crustal rocks. During the peak metamorphic event, the regional geothermal gradient was 30 to 40 °C/km, and the highest P, T, condition obtained from the lowest part of the granulite unit is ca. 830 °C, 7.0 kb. In this region, X_{H_2O} of Gt-Opx-Cd gneiss is about 0.15, and that of Gt-Cd-Bt gneiss is 0.4. The P-T condition of the granulite unit is well within a field where partial melting of pelitic and greywacke metamorphic rocks can take place by the fluid-present reaction of biotite and quartz with or without garnet (Vielzeuf and Holloway, 1988). This is evidenced by the restitic nature of Gt-Opx-Cd gneiss in the lowest part of the granulite unit.

The investigated results of P, T and X_{H_2O} of metamorphic rocks are significant for genetic considerations of the peraluminous granitic rocks in the Main Zone. The peraluminous granitic rocks are Opx-Gt-Bt tonalite (basal tonalite) in the granulite facies zone, Gt-Cd-Bt tonalite (lower tonalite) in the higher amphibolite facies zone and Cd-Mus-Bt tonalite (middle tonalite) in the lower amphibolite facies zone. Mineralogical and chemical natures of these strongly peraluminous tonalitic rocks, combined with the results on the metamorphic condition, permit us to regard these tonalites as derivatives from S-type granitic magma generated by crustal anatexis of pelitic metamorphic rocks in deeper crust.

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Appendix

Mineral abbreviations: And: andalusite, Bt: biotite, Cd: cordierite, Chl: chlorite, Co: corundum, Cpx: Ca-pyroxene, Cum: cummingtonite, Ep: epidote, Ged: gedrite, Gt: garnet, Hbl: hornblende, Hc: hercynite, Kfs: K-feldspar, Mus: muscovite, Opx: orthopyroxene, Phe: phengite, Pl: plagioclase, Qz: quartz, Sil: sillimanite, St: staurolite.

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