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INTERMEDIATE-P/T TYPE ARCHEAN METAMORPHISM OF THE ISUA SUPRACRUSTAL BELT: IMPLICATIONS FOR SECULAR CHANGE OF GEOTHERMAL GRADIENTS AT SUBDUCTION ZONES AND FOR ARCHEAN PLATE TECTONICS

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ABSTRACT. The Isua supracrustal belt (ISB) rocks are dated at about 3.8 Ga and constitute the oldest accretionary complex in the world. Petrochemical and geothermobarometric studies of over 1,500 rock samples in ISB enabled us to estimate the extent of regional metamorphism, petrotectonic environment and subduction- zone geothermal gradient in the Archean. The following lines of evidence indicate the first discovery of progressive, prograde metamorphism from greenschist (Zone A) through Ab-Ep-amphibolite (Zone B) to amphibolite facies (Zones C and D) in the northeast part of the Isua supracrustal belt: (1) systematic change of mineral paragenesis in metabasites and metapelites; (2) progressive change of composition of major metamorphic minerals, including plagioclase, amphibole, chlorite, epidote, and garnet; (3) normal zoning of amphibole and garnet; and (4) absence of any vestige of high-grade metamorphism even in the lowest metamorphic zone.

Geology and chronological constraints of ISB indicate that the regional metamorphism was related to the subduction of Archean lithosphere. Metamorphic pressures and temperatures of the metamorphism are estimated to be 5 to 7 kbar from Grt-Hbl-Pl-Qz geobarometry and 380° to 550°C from the Grt-Bt geothermometry in Zones B to D. These P-T estimates indicate that ISB was affected by progressive metamorphism of an intermediate P/T ratio metamorphic facies series, and that it records a much higher geothermal gradient of a subduction zone in the Archean than is known from the Phanerozoic. The high geothermal gradient may have resulted from the subduction of young lithosphere and a high potential temperature of mantle.

The Archean high geothermal gradient led to melting of thick oceanic crust in a thin oceanic plate, creating many huge granitic (tonalite, trondhjemite, and granodiorite) batholiths. The slab melting changed the oceanic crust (density = 3.07) into a denser Grt-bearing residue (density = 3.55), implying that TTG melt extraction provided a potential driving force for Archean plate tectonics. In addition to the preservation of the oldest accretionary complex, this suggests that Precambrian-type plate tectonics, whose driving force is slab-pull due to densification of the residue of oceanic crust as a consequence of slab melting, was already operating in the Early Archean. The transition from Precambrian-type to Phanerozoic-type plate tectonics may be caused by thinning of oceanic crust and thickening of oceanic lithosphere in the late Archean, due to decrease of mantle temperature.

INTRODUCTION

The 3.6 to 3.8 Ga Amîtsoq gneisses and their enclaves of supracrustal rocks (the Itsaq gneiss complex) have long been recognized as one of the oldest rock groups on Earth (see summary by Nutman and others, 1996). The Isukasia region is located in the northeast part of the Archean gneiss complex of southern West Greenland, and is composed dominantly of Amîtsoq orthogneiss and the Isua supracrustal belt (ISB). Despite recent discoveries of much older terrains or detrital mineral grains, the investigation of the ISB is still significant, because relatively low-grade metamorphism and deformation allow us to recognize original geological structures, including

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magmatic primary features such as pillow lava structures (Komiya and Maruyama, 1995; Appel and others, 1998; Komiya and others, 1999; Myers, 2001) and original sedimentary structures (Boak and Dymek, 1982; Nutman, 1986; Komiya and others, 1999; Fedo, 2000). Recent application of modern accretionary geology concepts to the ISB successfully showed that it is the oldest accretionary complex in the world. The presence of an early Archean imbricated accretionary prism indicates that modern-style plate tectonics has operated on Earth since the Early Archean (Komiya and others, 1999).

The number of metamorphic episodes within the Isua accretionary complex is still uncertain. In addition, it is still controversial whether the metamorphism was progressive or not. Boak and Dymek (1982) and Boak and others (1983) estimated the metamorphic P-T conditions from the paragenesis of Grt+St+Bt or Ky+Bt with excess phases of Qz±Pl±Ms±Ilm in Isua metapelites. Temperatures calculated from Fe-Mg partitioning between coexisting Grt and Bt in Grt-Bt-St pelitic gneiss indicate equilibration at $T \sim 550°C$ for Grt cores, and at $\sim 460°C$ for Grt rims. In addition, the presence of Ky indicates $P \sim 5$ kbar during the metamorphism. Although Fe/Mg ratios in St, Bt and Grt decrease systematically westward, Boak and Dymek concluded that the belt has undergone homogeneous middle amphibolite (Amp) facies grade metamorphism. In addition, in spite of the lack of any compelling evidence (Nutman, 1986), the main stage of amphibolite facies metamorphism was assumed to occur in the early Archean (Boak and Dymek, 1982). Nutman (1986) speculated that six thermal events resulted in repetition of metamorphism from early Archean to Proterozoic and that late Proterozoic metamorphism under greenschist (Gs) -facies condition resulted in an apparent progressive metamorphism in the ISB. Dymek and others (1988) estimated metamorphic temperatures up to 650°C from Al_2O_3 contents of Opx within meta-ultramafic rocks in seven localities. Mineral name abbreviations follow Kretz (1983).

Grambling (1981) postulated a secular variation of metamorphic facies series from a compilation of Archean metamorphism. However, without determination of the tectonic setting, it is not clear whether the apparent secular change of metamorphic facies results from the secular variation of the geothermal gradient in subduction zones or from a difference of tectonic setting of the metamorphism. Miyashiro (1973) considered that the metamorphic facies series strongly depends upon the tectonic setting of the metamorphism. The metamorphism of ISB is exceptional because its tectonic setting is well known (Friend and others, 1987; Komiya and others, 1999). We report here the results of a mineralogical and petrological study of metabasites and metapelites affected by Archean progressive metamorphism from the ISB, West Greenland. We use mineralogical data to place constraints on the physical conditions during metamorphism, and discuss the evidence for a secular change of geothermal gradients at subduction zones, and the implications for a potential driving force for Archean plate tectonics.

GEOLOGICAL OUTLINE

The Isua supracrustal belt (ISB) lies approximately 150 km northeast of Nuuk, Greenland (fig. 1). It forms a 35 km-long arcuate tract and is the largest fragment of early Archean supracrustal rocks in the Akulleq terrane. Further south, early Archean supracrustal rocks, some equivalent to the Isua supracrustal rocks, occur as enclaves within the Amîtsoq gneisses.

In spite of metamorphism, metasomatism and polyphase ductile deformation, the original structures of the protoliths are well preserved in the northeastern part of the ISB where the total strain was relatively low. Well preserved primary structures, such as pillows in lavas and graded bedding allow us to work out their precursors (Komiya and Maruyama, 1995; Appel and others, 1998; Komiya and others, 1999; Myers, 2001; Fedo, 2000). The ISB comprises the following five major lithologic units: (1) mafic to felsic

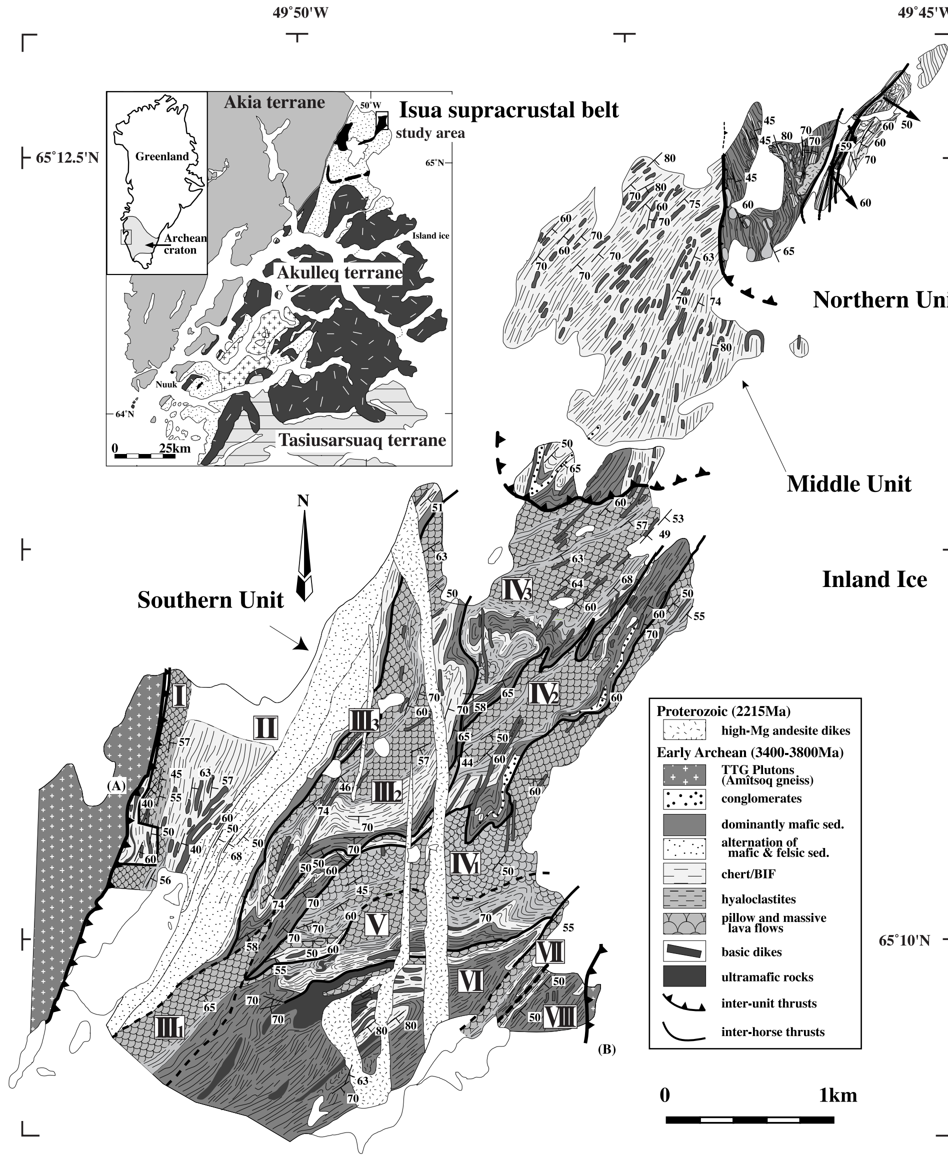

Fig. 1. Geologic map of the northeast part of the Isua supracrustal belt (Komiya and others, 1999). The inset is a geotectonic map of southern West Greenland, simplified after McGregor and others (1991). Three terranes, Akia (gray), Tasiusarsuaq (gray with horizontal bars), and Akulleq are shown. Akulleq terrane is composed of Ikkattoq gneiss (dark gray with white bars), Amîtsoq gneiss (stippled), and supracrustal rocks (black). Post-collision Qôrqut granite (cross) intrudes them. The Isua supracrustal belt is located in the northeastern corner.

turbidites, including interlayered minor conglomerate; (2) chert and banded iron-formation (hereafter BIF); (3) basaltic volcanics, including pillow lava, massive lava, pillow breccia, and hyaloclastite; (4) dolerite-gabbro intrusives related to (3); and (5) ultramafic rocks. In addition, there are many sills and dikes, some of which are mid-Archean Ameralik amphibolite dykes displaying characteristic large patches of plagioclase crystals, but the majority have, as yet, an unknown age. North south trending high Mg andesite dykes (HMA dykes) are the youngest rocks in the mapped area; they have a 2215 Ma zircon SHRIMP age (Nutman and others, 1995).

The northeastern part of the ISB is divided into three units bound by low-angle thrusts: the northern, middle, and southern units of figure 1. The southern unit, the best exposed, comprises fourteen thrust-bound horses sharing a similar lithostratigraphy, bound by at least ten layer-parallel thrusts. Duplex structures are widespread in the Isua belt, and vary from a few meters to kilometers in size. Duplexing proceeded from south to north, and is well documented in the relationship between link- and roof-thrusts. The reconstructed lithostratigraphy of each horse reveals a simple pattern, which consists of basaltic greenstone with/without pillow lava structure, chert/BIF and turbidite in ascending stratigraphic order. The cherts and underlying basalts do not contain continent- or arc-derived material. The lithostratigraphy is similar to Phanerozoic “oceanic plate stratigraphy”, except for the abundance of mafic material in the turbidites. The presence of “oceanic plate stratigraphy” and duplex structure indicates that the ISB is the oldest accretionary complex in the world. The predominant mafic composition of the turbidites suggests that they were deposited in an intra-oceanic environment, comparable to the present-day western Pacific Ocean (Komiya and others, 1999).

METAMORPHIC ZONATION

According to Komiya and others (1999), the ISB was affected by at least five thermal events: hydrothermal metamorphism, intrusion of tonalitic batholiths (precursor of Amîtsoq gneisses), Archean regional metamorphism, low-temperature mylonitization between Amîtsoq gneiss and ISB, and contact metamorphism caused by the intrusion of HMA dykes. Recently, Appel and others (2001) recognized a vestige of hydrothermal metamorphism from the cross-cutting relationship of a generation of quartz. In contrast, the intrusion of the HMA dykes caused significant low P/T contact metamorphism, producing mineral assemblages of Hbl+Opx+calcic Pl+Qz in metabasites, Ol+Opx+talc+Act-Hbl or Ol+Opx+Srp+Ath in ultramafic rocks, and Crd+And+Chl in metapelites; however, the thermal influence was limited to the immediate margins of the dykes (Hayashi and others, 2000). Mylonitization between Amîtsoq gneiss and ISB under greenschist facies conditions (Nutman, 1984; Komiya and others, 1999) affected the mineral assemblages and the composition of some minerals in Horse I and II (Komiya and others, 1999) or Sequence B (Nutman, 1986).

Metamorphic grade of the regional metamorphism is estimated from mineral assemblages and microprobe analyses of coexisting minerals in metabasites, metapelites, calc-silicate rocks and meta-ultramafic rocks. The area is divided into four metamorphic zones based on mineral assemblages in moderately magnesian metabasites, where chlorite is 50 to 60 in $Mg^{\#}$ ($Mg^{\#}=[Mg/(Mg + Fe^{T})]$ 100) (fig. 2). Zone A is defined by the Gs-facies mineral assemblage, Chl+Ep+Ab+Qz. Zone B is defined by the appearance of Ca-amphiboles, either Act and/or Hbl, together with Ab and/or oligoclase; the mineral assemblage is Chl+Ep+two Ca-amphiboles+peristeritic Pl+Qz, showing the transition between Gs and Amp facies. Zone C is defined by the appearance of calcic plagioclase, the appearance of a single amphibole of hornblende composition in magnesian samples, and the appearance of Grt coexisting with Hbl in iron-rich samples. The diagnostic mineral assemblage in metabasites is Chl+Ep±Act+Hbl+Pl+Qz. Zone D is defined by the disappearance of two Ca-amphiboles-bearing samples, concomitant with the high anorthite (An) content of plagioclase and a single Ca-amphibole of hornblende composition in moderately magnesian samples; the mineral assemblage is Chl+Ep+Hbl+calcic Pl+Qz. High magnesian rocks contain two amphiboles, cummingtonite (Cum) and Hbl.

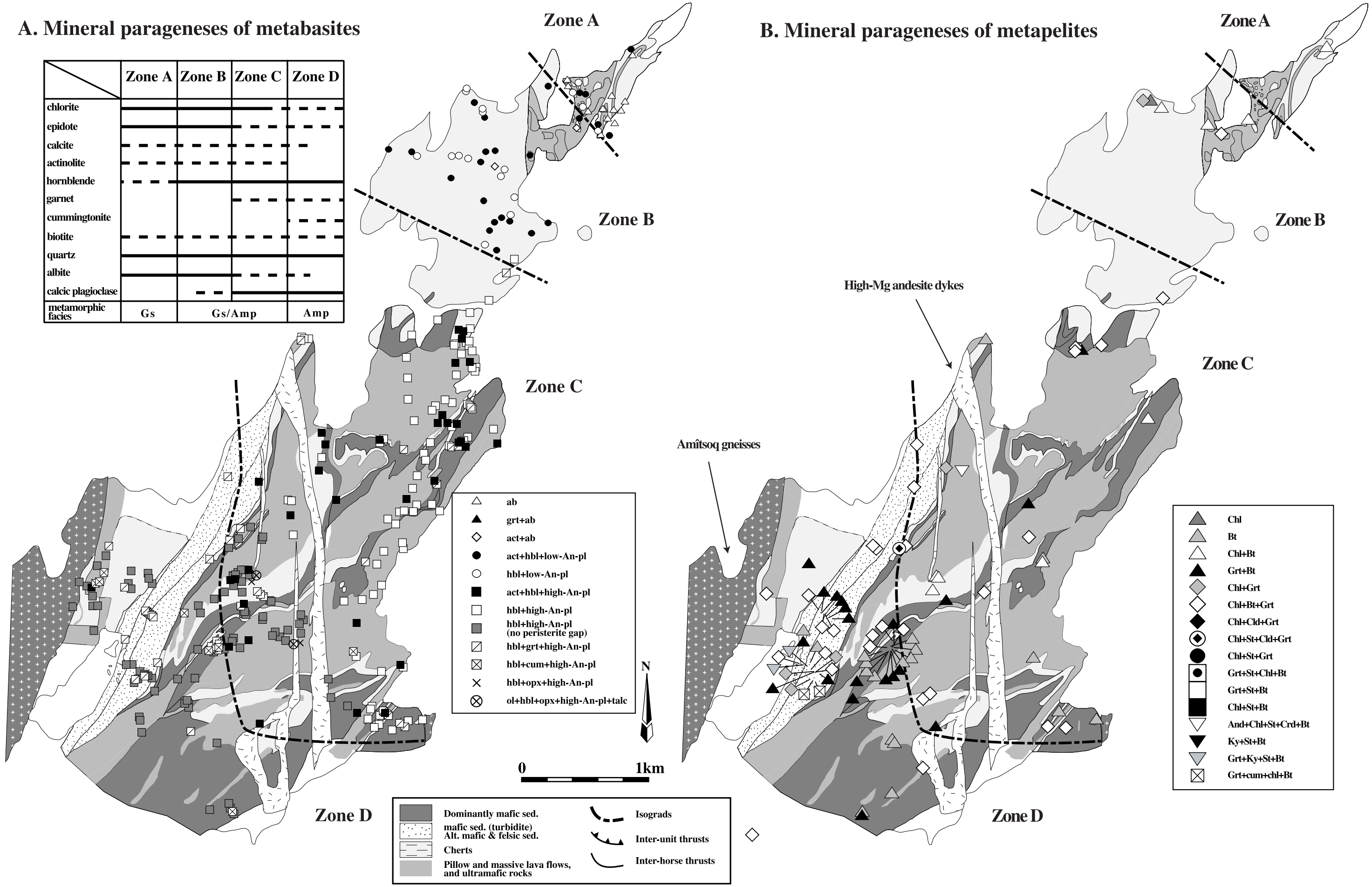

Fig. 2. Mineral assemblage of metabasites (A) and metapelites (B) in the northeastern area, which is subdivided into four zones based on mineral assemblages and anorthite (An) content of plagioclase in metabasite. The metamorphic grade ranges from greenschist facies in the northernmost area through epidote-amphibolite transition to amphibolite facies in the Southern Unit.

MINERALOGY

Mineral compositions were determined mainly by scanning electron microprobes (JEOL-733Mk.2 and JEOL JSM-840) with a Link EDS system at the University of Tokyo and an electron probe microanalyzer (JEOL-JXA-8800) at the Tokyo Institute of Technology. All analyses with the electron probe microanalyzers, except for Ni and Mn contents in olivine, were performed with an accelerating voltage of 15 kV, 12 nA beam current and a counting time of 10 to 40s. The analyses of Ni and Mn in olivine were performed with an accelerating voltage of 25kV, 30nA beam current and 50s counting time in order to obtain higher counts of secondary X-rays. The oxide ZAF correction method was employed. Analytical conditions of the scanning microprobe were 15 kV accelerating voltage, 1 nA specimen current and 60 to 80s counting time. Rare earth element (REE) abundances were obtained using a secondary ion mass spectrometer SIMS (Cameca Ims-3f) instrument at the Tokyo Institute of Technology. A well-calibrated augite megacryst from an alkali basalt in Japan was used as a standard for analyses of both pyroxene and metamorphic minerals. An energy-filtering technique with an offset voltage of -60 V for REE analyses was applied to eliminate possible molecular interferences. The primary ion beam $^{16}O^{-}$ was about 20 μm in diameter. Analytical uncertainties originate mainly from counting statistics and are 10 to 20 percent, the details of analyses were presented elsewhere (Wang and Yurimoto, 1994).

Metabasites

According to Komiya and others (1999), greenstones in the area were subdivided into MORB and OIB-types, based on their field occurrence. However, no differences between them exist in paragenesis and the composition of metamorphic minerals, indicating that metamorphism post-dated the emplacement of both MORB and OIB-type greenstones. Although severe recrystallization of Cpx and Pl into Ca-amphibole+Chl and Ab has obliterated original igneous textures in most samples, igneous variolitic and intersertal textures are preserved in some metabasites, even in Zone D (Hayashi and others, 2000). Moreover, relict cores of igneous calcic Cpx mantled by Chl and Ca-amphibole occur in a gabbroic rock (K025) of MOR-type in Zone D (Hayashi and others, 2000).

Plagioclase.—Plagioclase is ubiquitous in most metabasites. Some plagioclase is coarse-grained and exhibits cleavage and faint twinning. Where igneous textures are preserved in some metabasites, plagioclase forms interlocking laths with granular amphibole and chlorite.

Anorthite (An) contents are plotted on a frequency diagram of Zones across the area from northeast to southwest (fig. 3A). In Zone A Pl in six samples is albite except for four points, and its composition ranges between 0.4 and 22 in An content, 100 x Ca/(Ca+Na). Most Pl in Zone B is Ab and oligoclase, and a peristerite gap between $An_{7\sim18}$ and An_{20} is typically present. Pl forms composite grains with sharp interfaces. In Zone C an obvious peristerite gap is absent, except for a faint compositional gap in ILW4. Although Pl varies in An content in some samples, they are predominantly around An_{30}. In Zone D the composition of Pl becomes wider both within each sample and among samples. The An content tends to increase from Zone A to Zone D as metamorphic grade increases. In summary, the An content of Pl coexisting with Ca-amphibole+Ep+Chl+Qz in metabasites is used as an indicator of metamorphic grade, following the reaction

$$0.95\ \mathrm{Chl} + 0.93\ \mathrm{Ep} + 0.23\ \mathrm{Ab} + 0.18\ \mathrm{Qz} = 0.23\ \mathrm{Prg} + 1.41\ \mathrm{An} + 1.00\ \mathrm{H_2O} \quad (1)$$

Ca-amphibole.—The textural relationships of amphiboles, especially Act and Hbl, are divided into the following six types, based on microscopic observation and

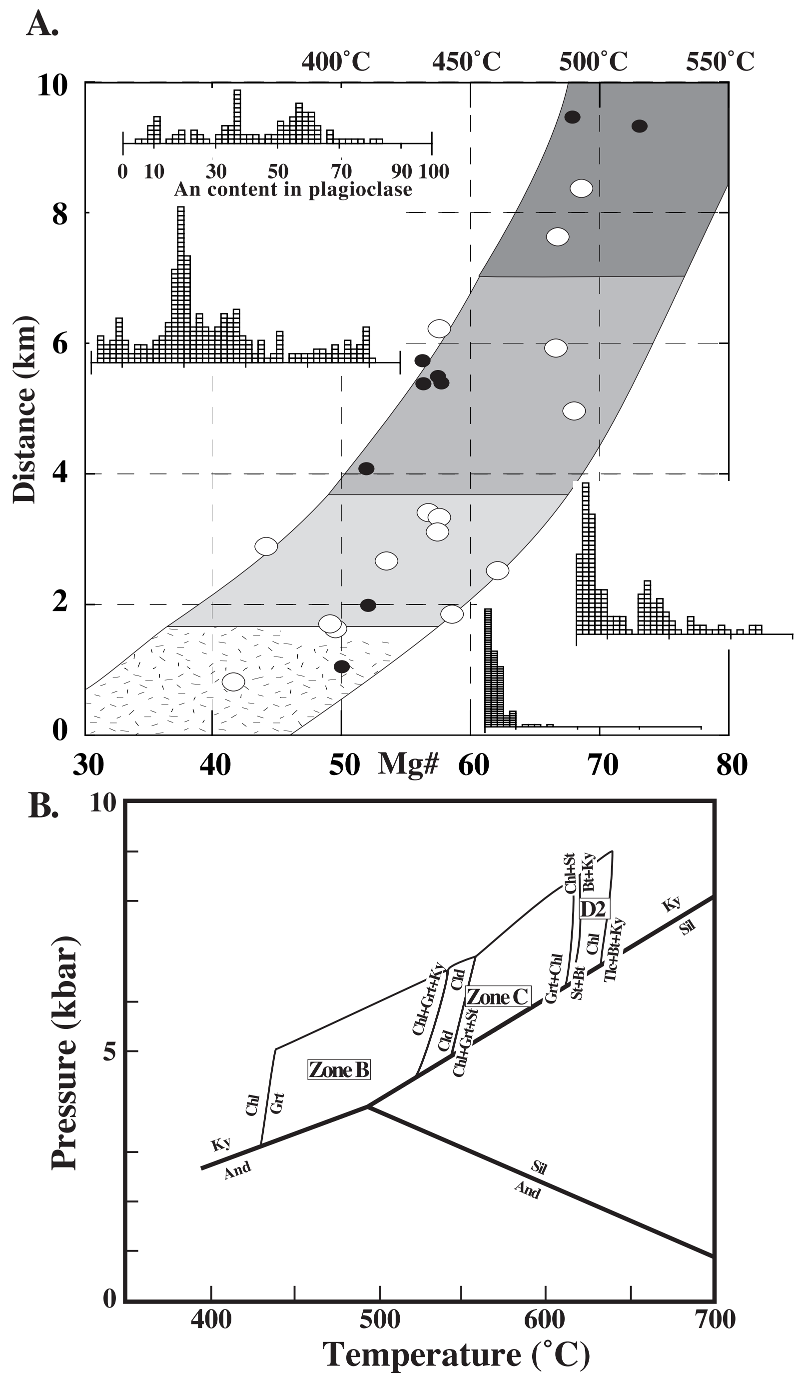

Fig. 3. (A) Frequency diagrams of An content of plagioclase in metabasites from Zone A to D. Relationship between Mg# of chlorite coexisting with Act, Hbl, Ep, Pl, Qz and opaque in metabasites from Zones B to D and distance from the northeastern edge of the Isua supracrustal belt (open circles). Relationship of metamorphic temperature versus distance (solid circles) is also shown. Metamorphic temperature is estimated from garnet-biotite geothermometry of Fe-Mg partitioning between garnet and biotite (Ferry and Spear, 1978). (B) P-T diagram inferred by mineral assemblage of metapelites (Spear, 1993). Stability field of garnet-bearing mineral assemblages, especially Chl+Grt, expands due to effect of whole-rock MnO content (Symmes and Ferry, 1992).

microprobe analyses: (1) only Act, (2) only Hbl, (3) composite grains of Hbl + Act without a compositional break, (4) composite grains of Hbl + Act with a sharp compositional break (fig. 4A), (5) Act mantled by Hbl without a compositional break, and (6) Act mantled by Hbl with a compositional break (fig. 4B). Types (3) and (6) are considered as the coexistence of two phases, but the others as a single amphibole phase. Especially, amphibole forms a continuous solid solution from Act to Hbl under high temperature and slightly high-pressure conditions, and this single amphibole phase is hereafter called actinolitic hornblende. The textures of types (5) and (6) are possibly caused by the overgrowth of amphiboles during prograde metamorphism, whereas that of type (4) reflects two amphibole phases in equilibrium.

The appearance of amphibole and its compositional change can be also explained by reaction (1). However, the stability field of amphiboles is influenced by pressure and bulk composition as well as temperature. Especially, the Fe/(Fe+Mg) ratio strongly affects the stability field. A higher Fe/(Fe+Mg) ratio leads to expansion of the stability field of both Act and Hbl and a wider compositional gap between Act and Hbl. Therefore, the composition and texture of amphiboles should be considered not only in relation to metamorphic grade but also to Fe/(Fe+Mg) ratios. This work adopted $Mg^{\#}$ of Chl coexisting with amphibole as an effective bulk composition instead $Mg^{\#}$ of the whole rock because of domain equilibrium.

This paper presents only a short summary of the relationship between the composition of amphibole and chlorite; detailed descriptions were given elsewhere (Hayashi and others, 2000). In Zone A, no amphibole exists in moderately magnesian samples, where the $Mg^{\#}$ of chlorite is 50 to 60. However, two amphiboles (Act and Hbl) are present in Fe-rich samples with $Mg^{\#}$ 42 to 50 of chlorite. In Zone B, two amphiboles are present in twelve metabasites where $Mg^{\#}$ of chlorite is in the range between 44 and 62 (fig. 5). A single amphibole of Hbl occurs in twenty highly Fe-rich metabasites, with less than 44 in chlorite-$Mg^{\#}$. Zone C is characterized by the presence of actinolitic hornblende in the highest chlorite-$Mg^{\#}$ samples. In the range 60 to 70 in $Mg^{\#}$ of Chl, relatively low magnesian samples contain two Ca-amphiboles, whereas others contain only Hbl. In the range between 50 and 60 in $Mg^{\#}$ of Chl, some moderately magnesian metabasites have two phases of Act and Hbl, whereas others have Hbl with/without Grt. In Zone D, composite grains of Act and Hbl of type (4) disappear in moderately magnesian metabasites, and instead, composite grains of Hbl and Cum are found in some samples (fig. 2A). The moderately magnesian samples have both Grt and Hbl. Actinolitic hornblende is present, where $Mg^{\#}$ of chlorites is more than 60.

In summary, $Mg^{\#}$ of metabasites, which contain both Act and Hbl, increases with increasing metamorphic grade from Zone A to Zone D (fig. 5). Moreover, metabasites, which have only Hbl with/without Grt, also become synchronously more magnesian. At a given $Mg^{\#}$ of about 55, the mineralogy of amphibole is well correlated with metamorphic grade: no amphibole in Zone A, presence of Act and Hbl of type (4) in Zone B, either only Hbl or two phases of Act and Hbl of type (4) in Zone C, Hbl with Grt in Zone D1, and actinolitic hornblende in Zone D2 (fig. 5). Provided that the metabasites in ISB can be expressed in the $NCF^{2+}Fe^{3+}MASH$ system, mineral compositions in the metabasites including Act (or Grt), Hbl, Pl, Qz, Ep, Chl, opaque minerals and H_2O fluid must be fixed under given P-T conditions, and change with metamorphic grade. Therefore, when $Mg^{\#}$ of chlorite coexists with Act+Hbl or with Grt+Hbl metamorphic temperature increases (fig. 3A). The relationship indicates progressive increase of metamorphic grade from Zones A to D.

Other minerals.—Garnet occurs together with Hbl in iron-rich metabasites in Zones C and D (fig. 2A). It is euhedral, pink to reddish brown, up to a few cm across, and irregularly fractured. Garnet is rich in Fe and Mn, and has a relatively narrow compositional range within each sample. However, substantial variation is found

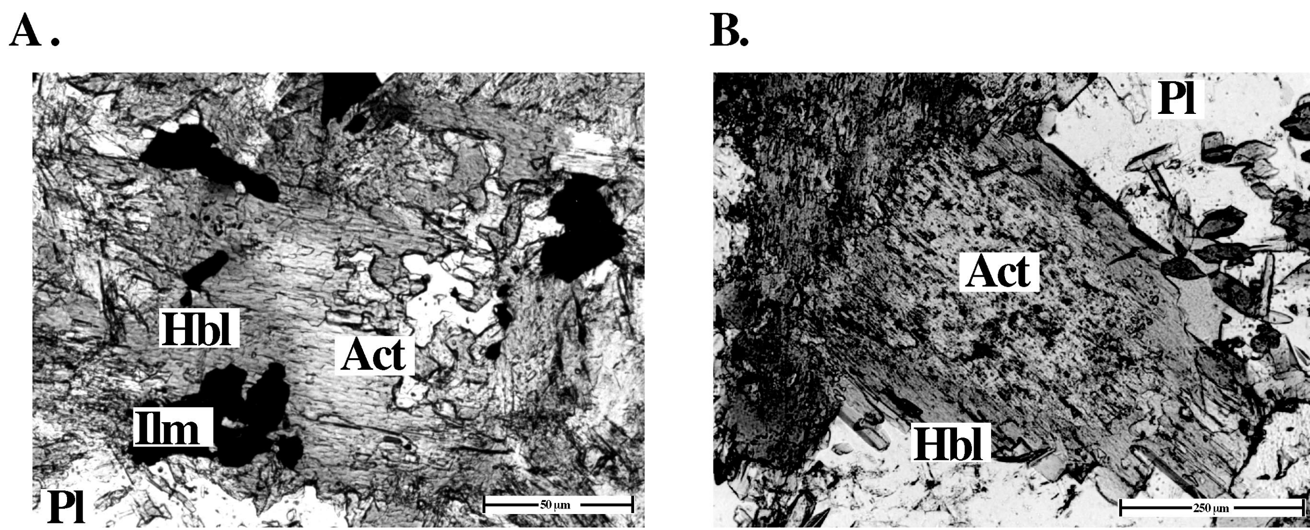

Fig. 4. (A) A composite grain of two Ca-amphiboles (Act and Hbl) with a sharp compositional boundary in K465. Both Act and Hbl occur as a mosaic throughout a composite grain. (B) Occurrence of two Ca-amphiboles in the high-grade amphibolite zone in IBFV09. Ca-amphiboles possess sharp zoning of actinolite in the core and hornblende in the rim, suggesting the presence and preservation of the prograde metamorphic stage.

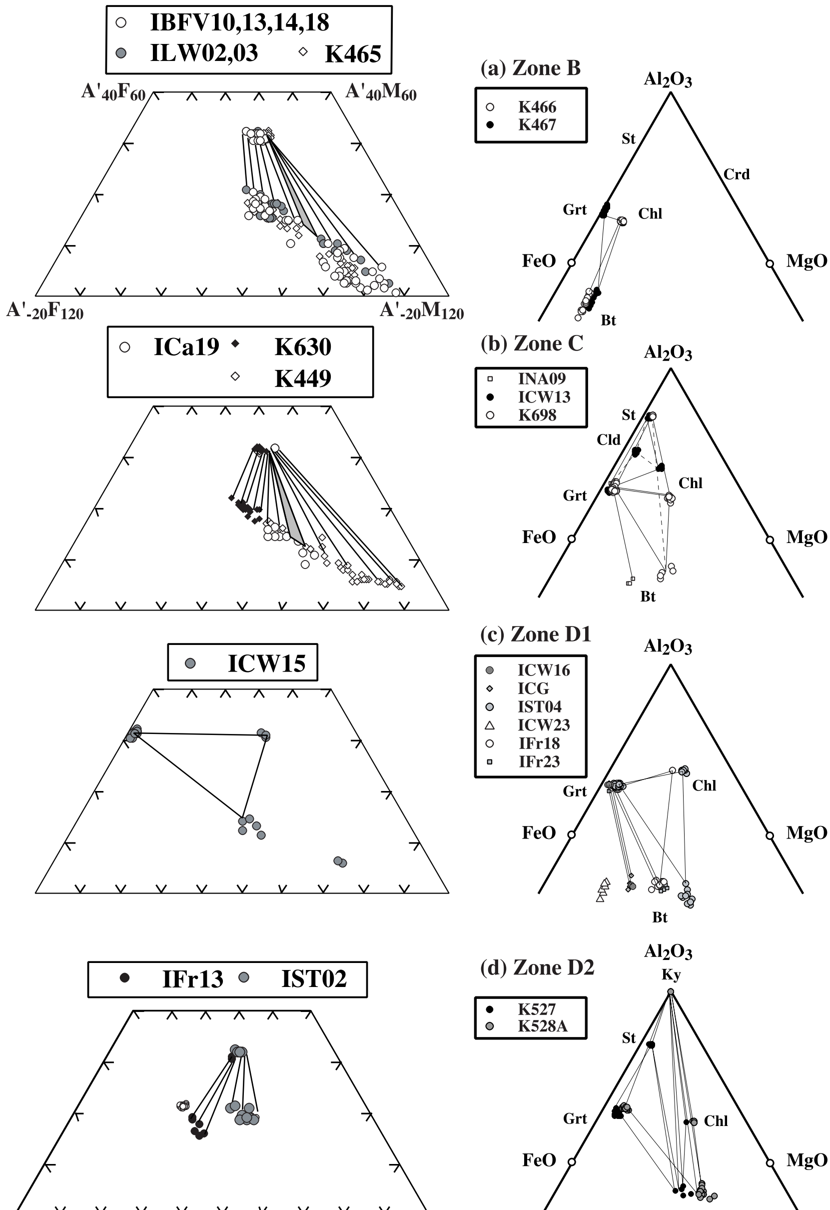

Fig. 5. (A) Mineral composition in metabasite containing chlorite with $Mg^{\#} = 50 \sim 60$ from Zones B to D onto A'FM plane projected from epidote (Harte and Graham, 1975). Excess phases are epidote, quartz, plagioclase and H_2O-rich fluid. (B) Compositions of minerals in metapelites from Zones B to D projected from quartz, muscovite, albite and H_2O on Thompson's AFM diagrams (Thompson, 1957).

among samples, suggesting the control of bulk rock composition on garnet composition. Garnet shows normal zoning of decreasing Mn, and of increasing Fe from core to rim, but no or faint chemical variation in Mg content. Ca content increases slightly from core to mantle, and then decreases from the mantle to the rim. Despite the large chemical variation among the samples, Ca content of garnet increases and Mn content decreases with increasing metamorphic grade (Hayashi and others, 2000). Chlorite occurs in all zones, but the modal amount decreases with increase of metamorphic grade. Large compositional variations are apparent within a single zone. However, within a single sample, chlorite is more homogeneous than coexisting Ca-amphiboles in $Mg^{\#}$ because compositional diffusion in chlorite is faster than in Ca-amphiboles. The $Mg^{\#}$ and the Si content (atoms per formula unit), on the basis of 28 oxygens, are in the range 40.8 to 57.5 and 5.3 to 5.7 in Zone A, 42.5 to 63.3 and 5.2 to 5.9 in Zone B, 42.4 to 77.7 and 5.1 to 6.1 in Zone C, and 54.0 to 77.4 and 5.1 to 6.1 in Zone D. In particular, the composition of Chl coexisting with Act+Hbl+Ep+Pl+Qz+opaque phase shows correlation with metamorphic grade (fig. 3A). Epidote is one of the most common Ca-Al silicates in all zones, but the modal amount decreases with increase of metamorphic grade. It forms tiny granules scattered throughout the groundmass, or as interspersed aggregates of anhedral epidote. Although most epidote grains are homogeneous, some epidotes in Zone A show chemical zoning of XFe^{3+} (Hayashi and others, 2000). The XFe^{3+} decreases from 0.17 to 0.35 in Zone A, through 0.05 to 0.30 in Zone B to 0.03 to 0.26 in Zone C, but does not further decrease to Zone D (0.08-0.23). Titanite is an accessory, but ubiquitous phase in all metabasites. It forms dusty aggregates of tiny granules or coarse-grained idiomorphic, elongate grains with compositions ($Al_{2.7\text{-}5.1}Ti_{91.6\text{-}96.1}Fe^{3+}{}_{1.2\text{-}3.3}$) in Zone A, ($Al_{2.6\text{-}7.5}Ti_{89.5\text{-}96.2}Fe^{3+}{}_{0.2\text{-}3.6}$) in Zone B, ($Al_{2.9\text{-}7.5}Ti_{91.1\text{-}96.6}Fe^{3+}{}_{0.4\text{-}1.9}$) in Zone C, and ($Al_{3.5\text{-}4.3}Ti_{94.6\text{-}95.4}Fe^{3+}{}_{1.0\text{-}1.1}$) in Zone D, comparable to compositions of titanite in epidote zone of Franciscan metabasite in Phanerozoic (Maruyama and Liou, 1988). Fe_2O_3 content of titanite tends to decrease with increasing metamorphic grade. Titanite becomes less common with increasing metamorphic grade, and instead, ilmenite becomes stable. Quartz is ubiquitous in all zones, but calcite is restricted to low-grade samples. Scapolite is a minor phase, substituting for plagioclase. Cummingtonite occurs in relatively high (Fe + Mg)/(Ca + Al) ratio rocks in Zone D and does not coexist with epidote. Spinel, magnetite, hematite and sulfides are rare. Fine-grained pinkish Opx is present along aureoles of HMA dykes, indicating that the contact metamorphism reached the pyroxene hornfels facies.

Metapelites

Mafic terrigenous sediments at the stratigraphic top of each horse are enriched in MgO and FeO, relative to Phanerozoic shales. This chemical difference places the mineral assemblage in the basaltic, instead of the pelitic, system. Therefore, only about 120 samples have mineral assemblages of typical pelitic systems; Grt, Chl, Bt, Ms, Qz, St, Pl, Ky, Cld, with subordinate Ca- and Mg-amphiboles, Tur, Cal (calcite), Zrn, and opaque minerals (fig. 2B). Mineral assemblages of many samples are either Chl+Bt or Grt+Bt, together with Ms and Qz. However, some samples have diagnostic mineral assemblages that enable estimates of metamorphic temperatures and pressures; the mineral parageneses are shown in Thompson's AFM diagrams (Spear, 1993) (fig. 5). In addition, andalusite and Crd are present in the aureoles of HMA dykes, consistent with the mineral assemblage in metabasites in the contact aureole of HMA dykes. Small pumpellyites, ($Mg^{\#}$ = 58.8~53.9, and Al content = 4.90~5.13 atoms per formula unit on the basis of 24.5 oxygens) are found in garnet in Horse II in Zone D, consistent to low-grade metamorphism during mylonitization between the Amîtsoq gneiss and ISB (Komiya and others, 1999).

In Zone A, Grt is absent and Chl and Bt are present together with phengite (Phn) and Qz. Grt becomes stable in the northeastern side of Zone B. The mineral parageneses of Chl+Grt or Chl+Bt+Grt are found in some metapelitic rocks, indicating the stable coexistence of Grt and Chl (fig. 2B). In Zone C, Cld (ICW13) and St (K698) occur in aluminous samples together with Grt, Chl, Qz, and Phn. The

Fe/(Fe+Mg) ratio of the coexisting phases in ICW13 is lower than that in K698 (fig. 5). Although the mineral assemblage of other Zone C rocks is the same as that in Zone B, Fe/(Fe+Mg) ratios of the coexisting minerals are obviously different, and become higher than those in Zone B. In Zone D, there are two types of mineral paragenesis: most samples contain the same mineral assemblage of Grt+Chl+Bt as that in Zone C, but the distinctive parageneses of Ky+St+Bt, Ky+Chl+Bt or St+Grt+Bt in K527 and Ky+Grt+Bt or Ky+Chl+Bt in K528A occur on the southern side of the Zone (figs. 2B and 5). In summary, five reactions are expected in the phase diagram of the KFMAS+MnO system (Symmes and Ferry, 1992; Spear, 1993), Chl1=Grt+Chl2 (about 440°C), Chl+Grt+Ky=Cld (about 530°C), Cld=St+Grt+Chl+H_2O (about 560°C), Grt+Chl=St+Bt+H_2O (about 610°C) and St+Chl=Ky+Bt+H_2O (about 620°C) (fig. 3B).

REE Composition of Metamorphic and Igneous Clinopyroxenes, and Related Metamorphic Minerals

Two types of Cpx occur in ISB (Hayashi and others, 2000). One is metamorphic Cpx whose composition is salite, $Wo_{48.5\text{-}50.3}En_{29.6\text{-}33.0}Fs_{17.2\text{-}21.6}$ (fig. 6), formed by the regional metamorphism in a calc-silicate rock (Isc13). Another is relict igneous Cpx derived from crystallization of MORB, and is augite, $Wo_{28.9\text{-}38.1}En_{39.1\text{-}56.7}Fs_{12.0\text{-}26.5}$ in composition (fig. 6). Abundances of REE in the salitic Cpx are below detection limit, indicating the abundances of less than tens part per trillion. On the other hand, Ep coexisting with Cpx has very high concentrations of REE, and shows a flat REE pattern with a positive Eu anomaly (fig. 6). The low abundance of REE in the Cpx is possibly because Ep crystallized before Cpx and scavenged all REE throughout the samples.

The relict igneous Cpx have moderate abundances of REE, and display slightly LREE-depleted patterns (fig. 6). REE content in the magma coexisting with Cpx can be calculated from the REE abundances in the Cpx, using an appropriate partition coefficient (Nicholls and Harris, 1980). The result indicates that the REE pattern calculated from the Cpx is slightly LREE-enriched pattern, and is consistent with the REE pattern in the highly evolved residual melt of MORB in ISB (Komiya and Maruyama, 1995). Rims of Cpx were replaced by amphibole, Chl and tiny Ep during the Isua metamorphism. A metamorphic amphibole, replacing Cpx, shows a highly LREE-depleted pattern (fig. 6). Chl contains no REE. Although the small size of scattered epidote prevents REE analyses, the Ep are considered as a host mineral as well as amphibole for REE released from Cpx as shown above. In the case that the metamorphic minerals began to crystallize simultaneously, REE also may be shared among the metamorphic minerals.

GEOTHERMOBAROMETRY

In metapelites from Zones B to D, both Grt and Bt are present, allowing us to estimate the metamorphic temperature from Fe-Mg partitioning between Grt and Bt (Ferry and Spear, 1978). Grt commonly show remarkable zoning; most have normal zoning, whereas some show reverse zoning. Although rare, few Grt shows oscillatory zoning. Grt-Bt geothermometry indicates a progressive increase of metamorphic temperature from 400°C in Zone B, through about 450°C near the boundary between Zone C and D, to about 550°C in Zone D with increasing distance from the northeastern edge of ISB (fig. 3A). In addition, the pressure of Isua regional metamorphism can be estimated to 5 to 7 kbar from Grt-Hbl-Pl-Qz geobarometry (Hayashi and others, 2000). Thus, the Isua regional metamorphism belongs to the intermediate-P/T type metamorphic facies series (Miyashiro, 1973).

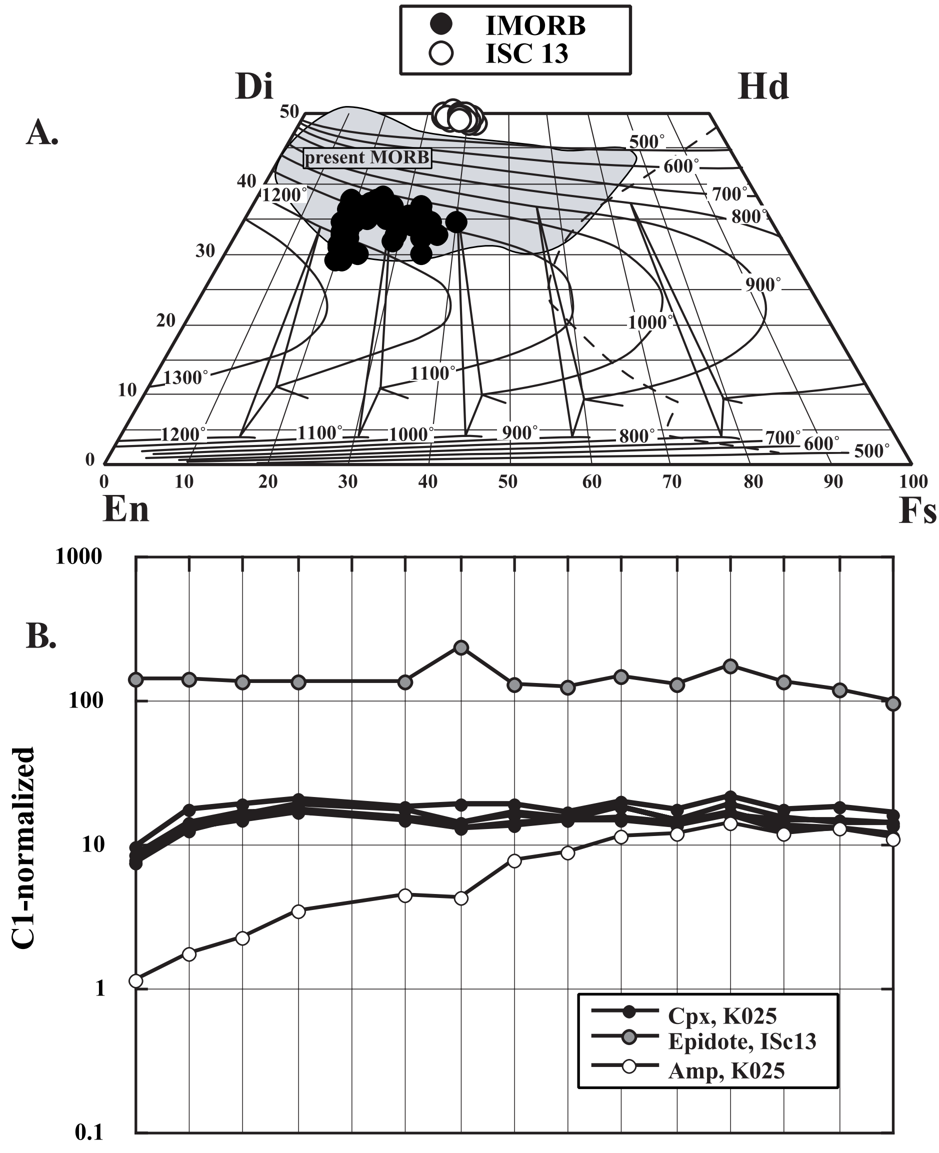

Fig. 6. (A) Compositional plots on Lindsley's quadrilateral diagrams (Lindsley, 1983) for relict igneous calcic clinopyroxene from a pillowed basalt in the Zone D and metamorphic clinopyroxene in calc-silicate rock (ISc13). (B) Chondrite-normalized rare earth element patterns of the relict igneous clinopyroxenes and metamorphic minerals of epidote and amphibole.

DISCUSSION

Thermal Structure of Isua Metamorphism, and Comparison with Archean Metamorphism Elsewhere

It had been long considered that ISB underwent a uniform regional metamorphism (Boak and Dymek, 1982). In spite of a slight increase in $Mg^{\#}$ of coexisting Grt and Bt from the northeast to the southwest, Boak and Dymek (1982) assumed a uniform regional metamorphism under amphibolite facies conditions. Nutman (1986) showed that Pl varies in An content throughout the ISB, but he presumed that the variation was caused by a subsequent alteration/metamorphism and that the main metamorphism in ISB was a uniform amphibolite facies metamorphism. However, this work presents the following evidence for an intermediate-P/T type progressive metamorphism in the northeastern area of the ISB: (1) systematic change of mineral parageneses in metabasites and metapelites; (2) progressive change of composition of major metamorphic minerals, including plagioclase, amphibole, chlorite, epidote, and garnet; (3) normal zoning of amphibole and garnet; and (4) absence of any vestige of high-grade metamorphism even in the lowest metamorphic zone.

The timing of the metamorphism of the Isua supracrustal belt is highly debatable as to whether the metamorphism occurred in the late or early Archean. U-Pb *in situ* zircon dating showed a 2.7 Ga-overgrowth of zircons in biotite quartzite, as evidence for late Archean metamorphism (Nutman and Collerson, 1991; Whitehouse and others, 1999; Nutman and others, 2000). In addition, U-Pb dating of multi-grain zircon (Baadsgaard and others, 1976; Baadsgaard, 1983), recent PbSL dating of magnetite in BIF (Frei and others, 1999), and a Sm-Nd isochron age of meta-basalts in western ISB (Frei and others, 2002) also indicate a late Archean isotopic disturbance. On the other hand, many previous works suggested that metamorphism and/or metasomatism occurred in the early Archean (Appel and others, 2001; Frei and Rosing, 2001; Frei and others, 2002). However, apparent similar mineral paragenesis in pillowed basalt and in white patches of an amphibolite dike (an Ameralik dike) suggests that the main progressive metamorphism occurred in the late Archean. The metamorphism was possibly caused by collision-amalgamation of the Akia and Akulleq terranes (Friend and others, 1987, 1996; Nutman and others, 1989; McGregor and others, 1991), equivalent to the collision-accretion of island arcs (Kimura, 1996) and the collision-amalgamation of (small) continental blocks (Maruyama and others, 1996) in the Phanerozoic. The metamorphism records the geothermal gradient at a subduction zone (Miyashiro, 1973). Therefore, we compare the metamorphism of the Isua supracrustal belt with collision-type metamorphism in the Phanerozoic (fig. 7).

All Archean metamorphic terrains worldwide belong to the intermediate- or low -P/T type metamorphic series (fig. 7; Grambling, 1981; Hayashi and others, 2000). We have subdivided 111 Archean metamorphic terrains into two groups; "metamorphic belt" and "metamorphic area," in "belts" the longer dimension is three times longer than the shorter side, whereas the aspect ratio of the "areas" is less than three. Although many Archean metamorphic terrains cannot be simply classified as either "belt" or "area", we have tentatively identified 20 "belts", 18 "areas" and an undifferentiated group. Ten terrains have a composite metamorphic structure, composed of different metamorphic units. All the >300 km long metamorphic belts including the longest metamorphic belt, the Stanovik fold, belong to the intermediate- P/T type metamorphic facies series. A low P/T-type metamorphic belt is relatively short, less than 600 km. Many Phanerozoic metamorphic belts extend for several thousands of kilometres along plate boundaries, whereas Archean metamorphic belts are much shorter, at most hundreds of km long. Such a difference may result from shorter plate boundaries, because of the presence of many microplates in the Archean. This idea is supported by numerical calculations of mantle convection under high temperature conditions (see Hargraves, 1986; de Wit and Hart, 1993).

The age of a subducting plate has a significant control on the geotherm (Martin, 1986; Defant and Drummond, 1990; Peacock, 1996; Hyndman and others, 1997; Harry and Green, 1999). Maruyama and others (1996) estimated a Phanerozoic geothermal gradient from the compilation of A-type (collisional) and B-type (Pacific) blueschist

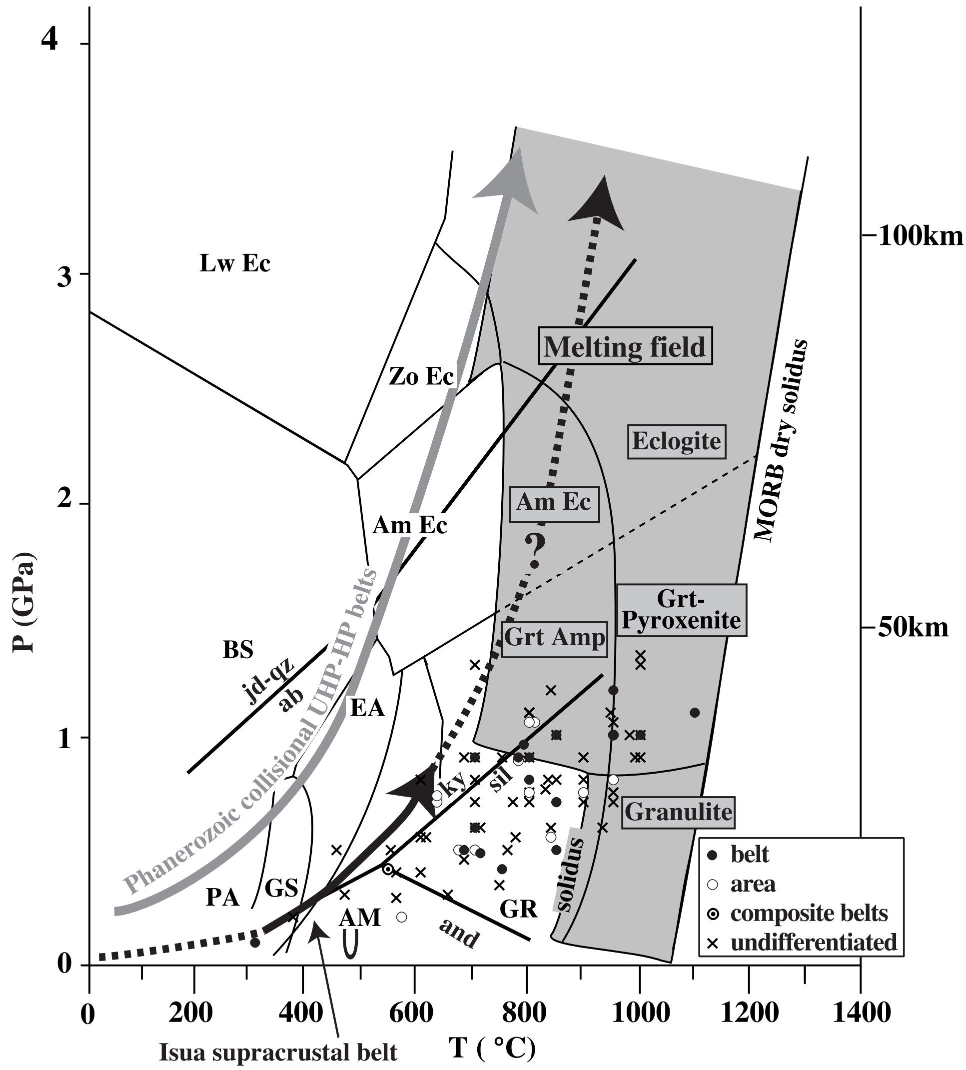

Fig. 7. Phase relations of hydrated MORB, modified after Okamoto and Maruyama (1999). The residual mineral assemblage (shaded region) after partial melting of hydrous Archean basalt is also shown (Rapp and Watson, 1995; Vielzeuf and Schmidt, 2001). Also shown are geothermal gradients of Phanerozoic collisional UHP-HP metamorphic belts (Maruyama and others, 1996) and Archean subduction zone metamorphism of the Isua supracrustal belt, accompanied by a recent compilation of P-T estimates of Archean metamorphic belts (Hayashi and others, 2000).

belts throughout the world, and this is reproduced in figure 7. Comparison of P-T estimates of Archean metamorphism obtained from the literature (Grambling, 1981; Hayashi and others, 2000) with Phanerozoic subduction metamorphism indicates that the geothermal gradient of Archean subduction zones was more than 200°C higher than in the Phanerozoic (fig. 7). The Phanerozoic geotherm ranges from low to high P/T-type, depending on the age of the subducted slab (Defant and Drummond, 1990; Peacock, 1996; Hyndman and others, 1997; Harry and Green, 1999), but the Archean geotherm does not extend to the high P/T side, ranging from low to intermediate

P/T-type (fig. 7). As mentioned above, determination of the tectonic setting of metamorphism is essential for the investigation of the secular change of metamorphism, because the metamorphic facies series strongly depends on the tectonic setting of the metamorphism (see Miyashiro, 1973). Our work indicates that the Isua progressive metamorphism belongs to the intermediate P/T-type metamorphic facies series. Significantly, the Limpopo metamorphic belt, which is a late Archean collision-type orogen between the Kaapvaal and the Zimbabwe cratons, also belongs to the intermediate P/T-type metamorphic facies series (see for example Tsunogae and others, 1992). These lines of evidence suggest that the geothermal gradient in Archean subduction zones was the intermediate P/T-type.

The following three candidates can be considered as a heat source for subduction zone metamorphism with a high geothermal gradient: (1) the overlying mantle wedge and continental crust, (2) the mantle below the oceanic lithosphere, and (3) the remaining heat of MORB-formation within the lithosphere. We have calculated the thermal structure of oceanic lithosphere, in order to verify whether the remaining heat of MORB formation within the lithosphere is possible as a heat source for Archean subduction metamorphism with a high geothermal gradient. The minimum temperature at the base of subducting oceanic crust can be calculated, based on the transient half-space cooling model and diffusivity of $1 mm^2 s^{-1}$ (Turcotte and Schubert, 1982). In the Phanerozoic, the temperature at the base of a 7 km thick oceanic crust falls below the solidus temperature of hydrated basalt only in 3 my after formation at given temperatures of the mantle (1330°C; Iwamori and others, 1995) and at the surface (0°C). However, in the case of Archean oceanic lithosphere with a 20 km thick oceanic crust (Ohta and others, 1996), the temperature at the base of the 20 km thick oceanic crust is still higher than the wet solidus even after 15 my at given temperatures of the mantle (1450°C; Ohta and others, 1996) and at the surface (0°C). This calculation indicates that the remaining heat of MORB formation within the lithosphere is a probable heat source for Archean metamorphism with a high geotherm, and even for slab melting.

A Driving Force for Archean Plate Tectonics

Recent applications of accretionary geology concepts to Archean greenstone belts have revealed that modern-style plate tectonics were already in operation in the Early Archean (Komiya and others, 1999). However, geophysical calculation of the relationship between the temperature of the mantle and the density of an oceanic plate contradicts the geological evidence for the operation of plate tectonics in the Early Archean, because an Archean oceanic plate capped by a thick oceanic crust may have been lighter than the asthenospheric mantle (Davies, 1992). The paradox is still unsolved, and must be evaluated.

Recent petrological study of Archean MORB indicated that the thickness of oceanic crust exceeded 17 to 18 km (Ohta and others, 1996), confirming earlier calculations of about 20 km (Sleep and Windley, 1982). On the other hand, Archean oceanic lithospheric plates were probably thinner than the Phanerozoic plates (despite the thicker Archean oceanic crus;t), because of the high temperature of the Archean mantle (Hoffman and Ranalli, 1988; Komiya and others, 1999). In the Phanerozoic, oceanic lithosphere is heavier than asthenosphere since its formation at a mid-oceanic ridge, because of negligibly thin oceanic crust (Kono and Yoshii, 1975). Therefore, even if the basaltic crust does not transform into eclogite during subduction, the oceanic lithosphere can subduct into the mantle due to its negative buoyancy. However, in the Archean, it is suggested that oceanic lithosphere could not directly subduct into the asthenosphere because it was lighter than the asthenosphere (Davies, 1992). However, even in this case, it is possible that the oceanic lithosphere reached beneath continental and younger oceanic lithosphere because relatively older oceanic lithosphere is denser than continental and relatively younger lithosphere.

The P-T path of the progressive metamorphism of ISB (this work), and the estimated P and T of worldwide Archean metamorphism (Grambling, 1981; Hayashi and others, 2000) are shown in figure 7. Although the Isua metamorphism did not reach such a high-grade condition up to the migmatite grade, the simple extension of the P-T path to the higher grade indicates that the hydrated oceanic crust was partially melted to produce the protolith of Amîtsoq gneiss. Namely, in the Archean, when the thickness of continental and older oceanic lithosphere exceeded approximately 30 km, the high geothermal gradient at the subduction zone led to melting of hydrated oceanic crust to form the tonalite-trondhjemite-granodiorite suite (TTG) (Martin, 1986; Defant and Drummond, 1990) and a denser garnet-bearing residue, such as high-pressure granulite (HGR) and garnet-clinopyroxenite. Assuming that modal and chemical compositions of the minerals in the garnet-clinopyroxenite and HGR follow melting experiments of Archean greenstones (Rapp and Watson, 1995), bulk densities of the garnet-clinopyroxenite and the HGR are calculated based on crystallographic data of Grt, Cpx, Opx and Pl (Smyth and McCormick, 1995). Densities of basaltic oceanic crust, and lithospheric and ambient mantles are calculated from parameters used by Davies (1992), for a case that the potential temperature of mantle is assumed to be 1480°C (Ohta and others, 1996). The calculated densities of basaltic crust, garnet-clinopyroxenite, HGR, and lithospheric and ambient mantle are shown in figure 8A. The calculation indicates that the average density of lithosphere increases from 3.22 to 3.46 for oceanic lithosphere with garnet-clinopyroxenite to 3.42 for oceanic lithosphere with HGR slab-pull force due to TTG melt extraction (fig. 8A). The calculation suggests that the transformation of basaltic crust to garnet-clinopyroxenite or to HGR was a potential driving force for Archean plate tectonics.

The Secular Change of Mode of Plate Tectonics

Some pioneering geologists suggested immediately after the establishment of the theory of the plate tectonics (see for example, Burke and others, 1976), that plate tectonics started on the Earth in the Archean, based on geological similarities with Phanerozoic orogenic belts. However, others suggested that plate tectonics had not operated until the late Proterozoic by reason of the absence of (1) ophiolites, (2) Franciscan-type mélange and (3) blueschists. Geological evidence for plate tectonics has been reported in some Archean terrains since the end of 1980's (Kusky, 1989; Hoffman 1991; Calvert and others, 1995). Recently, geological investigation of ISB revealed that it is the oldest accretionary complex, and that modern style-plate tectonics had been already in operation since the early Archean (see for example, Komiya and others, 1999). However, there are some differences between plate tectonic parameters in the Archean and the present (fig. 8). They are (1) thickness of oceanic lithosphere (Hoffman and Ranalli, 1988; Komiya and others, 1999), (2) thickness of the accompanied oceanic crust (Sleep and Windley, 1982; Ohta and others, 1996), (3) average life span of oceanic lithosphere (Hargraves, 1986; Komiya and Maruyama, 2002), and (4) the cause of slab-pull (this work).

In the Phanerozoic slab-pull is effective immediately after the birth of oceanic lithosphere, whereas in the Archean slab-pull is achievable only after slab melting. A ratio of the oceanic crust over the thickness of the lithosphere of about 0.75 is the critical number determining whether slab-pull is effective even just after the birth. Geophysical calculation of mantle temperature suggests that the ratio attained the critical value at about 1 Ga (Davies, 1992). In addition, compilation of modern and ancient ophiolites also supports the idea that thickness of oceanic crust decreased from about 12 to 5 km around 1 Ga (Moores, 1993). This evidence possibly indicates that Phanerozoic-type oceanic lithosphere with a slab-pull force from its birth began to be

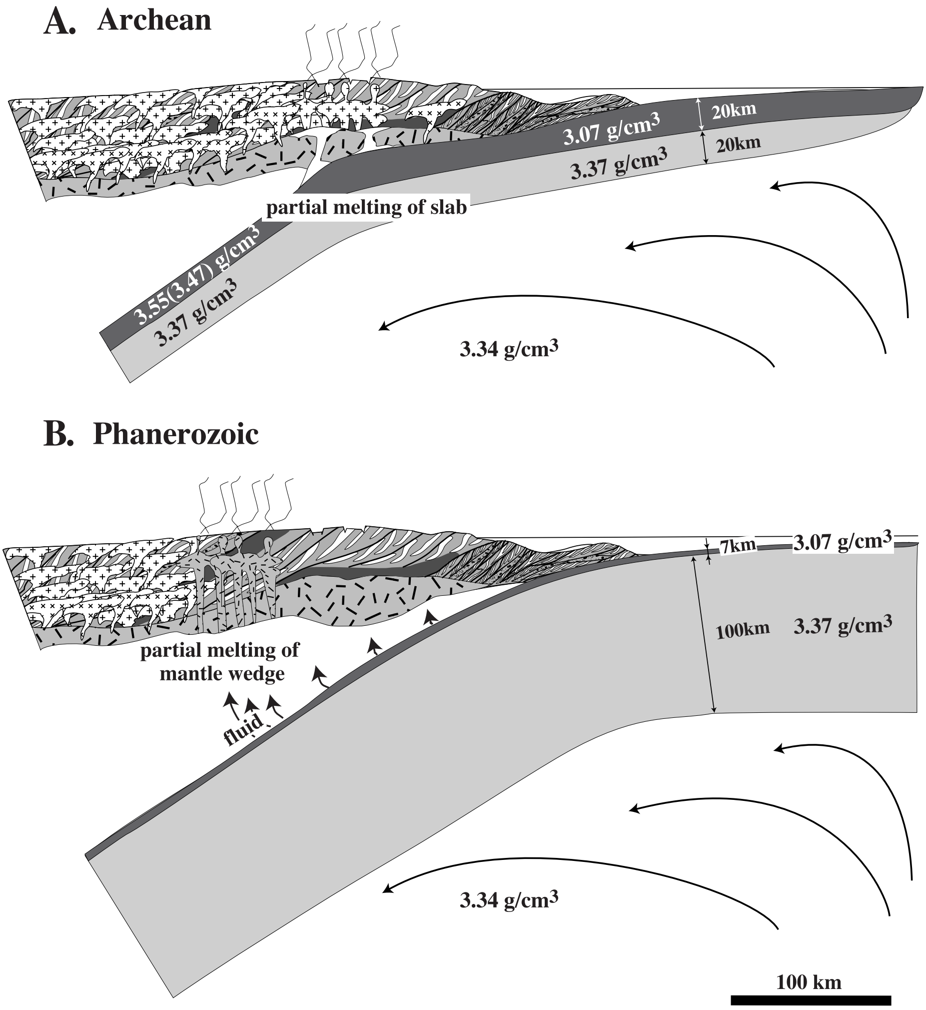

Fig. 8. Schematic illustration of subduction zones, showing contrasting tectonic modes of subducting lithosphere. (A) Low-angle subduction of young oceanic lithosphere results in slab melting, supplying felsic magma to the very small volume of mantle wedge (Precambrian-type). Calculated density profile of Archean oceanic lithosphere after TTG formation by slab melting indicates highly dense garnet-bearing residues after TTG melt extraction would become a potential slab-pull force. (B) High-angle subduction of old oceanic lithosphere supplies dehydration fluids to the hanging wall of the mantle wedge (Phanerozoic-type).

formed at about 1 Ga. In addition, blueschists did not form over the world before late Proterozoic (Maruyama and others, 1996). The presence of blueschists indicates that the geothermal gradient decreased at subduction zones, in other words, old oceanic plates began to subduct at that time. This line of evidence suggests that plate tectonics switched from a first-stage with slab melting (Precambrian-type) to a second-stage with/without slab melting (the Phanerozoic-type) in the late Proterozoic. In response to the initiation of Phanerozoic-type plate tectonics, mantle wedges increased in volume by the change from flat- to high-angle subduction, and simultaneously came to be metasomatized by H_2O-rich fluids (fig. 8B).

CONCLUSIONS

This work documents the following evidence for one of the oldest progressive, prograde metamorphisms from Greenschist through Epidote-Amphibolite to Amphibolite-facies in the northeastern area of the ISB. Metamorphic pressures and temperatures are estimated to be 5 to 7 kbar from Grt-Hbl-Pl-Qz geobarometry, 380° to 550°C from Grt-Bt geothermometry in Zones B to D, indicating an intermediate P/T ratio metamorphic facies series.

The Archean geothermal gradient in subduction zones, estimated from the progressive metamorphism of ISB, was much higher than the equivalent geotherm in the Phanerozoic. The high geothermal gradient may be caused by subduction of young oceanic lithosphere and high potential mantle temperature. The Archean high geothermal gradient led to melting of hydrated oceanic crust, which changed the oceanic crust to a denser garnet-bearing residue. The slab-pull caused by the slab melting was possibly the driving force of Archean plate tectonics.

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